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EVAPORATION

by

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ABSTRACT

This report provides an up to date review of current thinking concerning evaporation. It is divided into three sections:

I a resume of the physical theory involved in the natural evaporation process

II an outline of the methods used in evaporation measurement

III a survey and classification of the techniques used to estimate evaporation on the basis of its relationship to other measured variables.

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CONTENTS

I THE EVAPORATION PROCESS

Evaporation occurs when liquid water is converted into water vapour and transferred in this form into the atmosphere. The process can only occur naturally if there is an input of energy either from the sun or from the atmosphere itself, and is controlled by the rate at which the energy, in the form of water vapour, can diffuse away from the earth's surface. This diffusion occurs by the two rather different physical processes of molecular diffusion and turbulent diffusion at subsequent stages in the transfer path, but both processes are similar in that they can conveniently be represented by an analogy with electrical resistance. In this way the process of evaporation from a natural surface can be expressed on a physical basis by models which describe the effect of molecular and turbulent diffusion resistances on the partition of energy from the sun or the atmosphere. In Section I we outline the basic physical concepts and how they are combined to form a physical description. Section II describes the various techniques available for measuring evaporation, while Section III draws on the physical basis established earlier to classify and describe the many techniques available for estimating evaporation on the basis of measurements of related parameters.

1. ENERGY

Latent Heat

The molecules making up a volume of liquid water are in close proximity, with a separation of just over one molecular diameter. At such distances the subatomic particles, from which the molecules are built, interact in such a way that the molecules attract each other and there is a short range force between them. This force falls off rapidly as the separation increases, with the general shape illustrated schematically in Figure 1.

In water vapour the molecules are very much further apart, typically ten or more molecular diameters, depending on the vapour pressure. At this separation the intermolecular force is very small indeed. In order to create water vapour from liquid water it is necessary that the separation between all the molecules should increase. To do this it is necessary to do work against the force holding them together, that is to supply energy. The amount of energy required is of course directly related to the number of molecules, which is in turn directly proportional to the mass of water involved. The amount of energy per unit mass of liquid water is called the latent heat of vapourization of water, λ, and is 2.47×10^6 J kg at 10°C. It changes slightly with temperature, by about 0.1% per $^{\circ}$C, because the initial separation of the molecules making up the liquid varies with temperature (Figure 1).

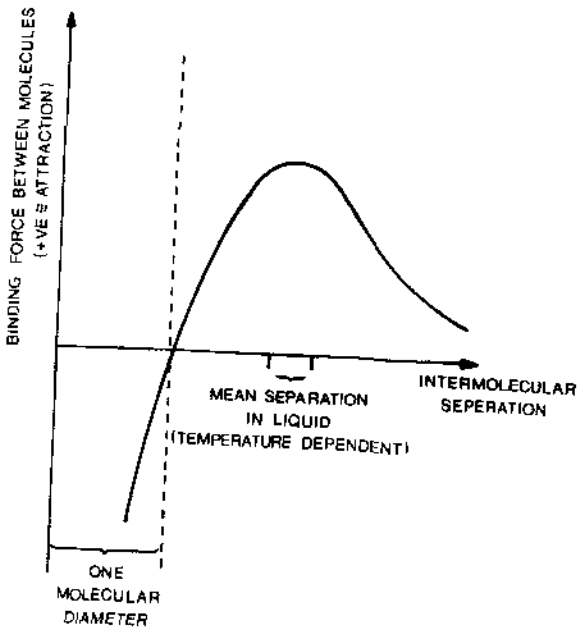

FIGURE 1

SCHEMATIC DIAGRAM OF THE ATTRACTIVE FORCE BETWEEN WATER MOLECULES AS A FUNCTION OF THE SEPARATION BETWEEN THEM

The flow of water vapour molecules away from an evaporating water surface necessarily implies a transfer of energy away from the surface in the form of latent heat. The amount of energy transferred is numerically equal to the product of the mass flow *i.e.* the evaporation, E, in kg. m^{-2} s^{-1}, and the latent heat of vapourization. In the following work we represent this energy flow by λE.

Liquid/Vapour Transfer at a Water Surface

Water vapour arises by the exchange of molecules with a free water surface somewhere in the system of vegetation and soil which makes up the interface between the earth and the atmosphere. This free water surface can either be inside the vegetation or soil (in 'dry' conditions) or on the surface of the vegetation or soil during and just after rainfall. It is useful to understand in outline the details of the physics involved at such a free water surface interface: for a more detailed description see Shuttleworth (1975).

Figure 2 shows schematically the exchange of individual water molecules at the boundary between the liquid and the vapour. The energy of the

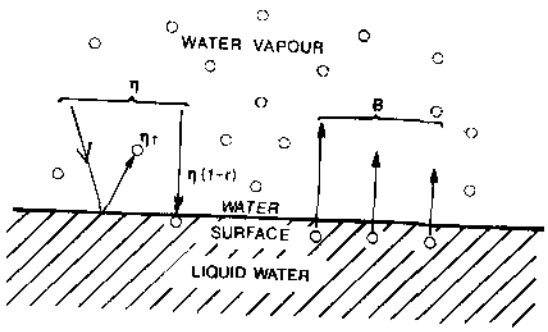

FIGURE 2

SCHEMATIC DIAGRAM OF THE EXCHANGE OF WATER MOLECULES BETWEEN LIQUID WATER AND WATER VAPOUR. B IS THE RATE AT WHICH MOLECULES 'BOIL OFF' FROM THE SURFACE: η IS THE RATE AT WHICH THEY COLLIDE WITH THE SURFACE; A FRACTION r ARE REFLECTED ON COLLISION

water molecules at the surface of a liquid follows a statistical distribution which is dictated by the surface temperature of the liquid, T_s. The number of molecules with energy greater than a particular value ε is given by

$$N_\varepsilon = \exp\left(-\frac{\varepsilon}{k'T_s}\right)$$

where T_s is *in degrees absolute* and k' is a constant called Boltzman's constant. To leave the surface a molecule must have an energy greater than (λ/n) when n is the number of molecules per unit volume of liquid water. The rate, B, at which molecules 'boil off' from the surface is therefore directly related to $N_{(\lambda/n)}$ *i.e.*

$$B = k_1 \exp\left(-\frac{\lambda/n}{kT_s}\right)$$

where k_1 is a constant.

At the same time molecules in the water vapour above the liquid approach the surface at a rate η, which is directly related to the vapour pressure, e, of the water vapour in contact with the surface. A certain fraction, r, of these vapour molecules are immediately reflected on collision with the surface, while the remaining fraction (1 - r) are absorbed. The number of molecules absorbed per unit time is therefore given by

$$\eta(1 - r) = k_2(1 - r)e$$

where k_2 is a second constant. The net evaporation rate, E, is therefore given by the difference between these two rates *i.e.* by

$$E = \left\{k_1 \exp\left(-\frac{\lambda/n}{kT_s}\right)\right\} - \left\{k_2(1 - r)e\right\} \qquad (1)$$

Saturation

In the previous section it was demonstrated that the evaporation from a free water surface is the difference between two rates one of which is determined by (absolute) temperature and the other by the vapour pressure (or concentration) of the water vapour above the surface. If molecules are able to diffuse freely away from the surface then the vapour pressure, e, adjacent to the surface can remain low, and the difference between these terms finite. In this situation it is possible to maintain a persistent flow of molecules from the liquid to the vapour, with an equivalent flow of latent heat.

If on the other hand, the volume of air above the liquid is sealed off, then it is no longer possible for the water molecules to diffuse freely away from the surface. As more molecules leave the surface the

concentration of the water vapour, and its equivalent vapour pressure, increases until such time as the two rates in equation (1) are equal and there is no longer any evaporation (E = 0). The sealed volume of air is then said to be 'saturated' and cannot absorb any more water molecules. At a given temperature this situation occurs at a particular vapour pressure, which is called the 'saturated vapour pressure', e_s. Substituting the condition E = 0 in equation (1) it is easily seen that

$$e_s = \frac{k_1}{k_2 (1 - r)} \exp \left(- \frac{\lambda/n}{kT_s} \right)$$

so that this vapour pressure is a unique function of T_s which can be measured experimentally.

Figure 3 shows the variation of saturated vapour pressure, e_s, as a function of temperature. It is an important fact in building physical models of evaporation, that not only is the variation of e_s a well measured function of temperature but so is its gradient $\left(\frac{\partial e_s}{\partial T} \right)$. This gradient, or more particularly Δ, which is the mean gradient between two temperatures T_1 and T_2,

$$\Delta = \frac{e_s(T_2) - e_s(T_1)}{T_2 - T_1} \qquad (2)$$

recurs frequently in equations describing evaporation, as we shall see later in the text.

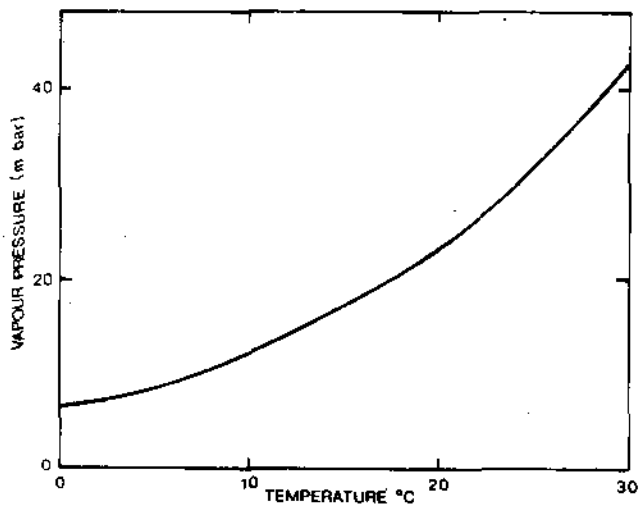

FIGURE 3

VARIATION OF SATURATED VAPOUR PRESSURE WITH TEMPERATURE

Sensible Heat

There are of course several ways in which the energy input to the earth's surface can be used apart from supporting evaporation. In practice

large quantities of energy are exchanged between the surface and the atmosphere by the joint processes of thermal conduction and thermal convection through the atmosphere. Thermal conduction only occurs over limited distances very close to the surface at which the radiant energy is intercepted; we return to this later in the text. Further away from the surface the primary exchange mechanism is thermal convection; convection occurs when mass motions of the air result in the transport and mixing of the heat content of the air between different levels in the atmosphere. We speak of the flow of energy between the surface and the atmosphere by the combined processes of thermal conduction and thermal convection as 'sensible' heat flux, because it is this transfer which determines air temperature, a property of the air that we can sense.

The flux of sensible heat is very commonly away from the surface in daytime conditions, when the temperature of the ground is usually greater than that of the atmosphere, but is usually towards the surface at night, when it supports the outward flux of radiation. If the surface of the vegetation is wet the evaporation rate is often very high, particularly for tall vegetation, and even during the day can exceed the energy reaching the surface as radiation. In such situations there will be an inward flux of sensible heat to provide the additional energy requirement.

Radiation

The sun provides the primary source of radiant energy input at the earth surface. The radiation input it provides is equivalent to that of a full radiator of about 6000 ^{o}K. The spectrum is modified by absorption in atmospheric gases, particularly water vapour; and by scattering from the individual molecules of the atmosphere in clear sky conditions, and additionally from clouds when these are present. Most of the radiation is confined to short wavelengths in the band 0.3 to 3 μm, the particular spectrum depending on the fraction of the total short wave energy input, S_T, reaching the ground in the direct solar beam. A significant part of the short wave energy, S_D, reaches the ground in a diffuse or non-directional form after scattering. The fraction varies but is typically 15 to 25% in clear sky conditions, approaching 100% in overcast conditions.

On reaching the ground part of the short wave radiation (with wavelength less than 3 μm) is reflected by the surface. The fraction reflected, the reflection coefficient or *albedo*, depends on the angle of incidence of the solar beam and the type of vegetation. The daily mean reflection coefficient is commonly 0.25 $\pm$ 0.05 for a broad range of short, agricultural crops but is more typically 0.15 $\pm$ 0.05 for taller crops such as tobacco, sugar cane and natural or man-made forest. This difference arises because tall vegetation is more able to absorb the solar beam by multiple reflections within the canopy: it is a point which should be remembered when estimating evaporation for different crop types.

In addition to the energy received directly and indirectly from the sun, there is a significant exchange of radiant energy between the earth's surface and the atmosphere in the form of radiation at longer wavelengths, in the range 3 - 100 μm. This radiation arises because both the surface and the atmosphere emit 'black body' radiation with a spectrum characteristic of their temperature in degrees absolute. The measured spectrum of incoming longwave radiation near the surface is greatly modified by the absorption by the atmospheric gases, particularly water vapour and carbon dioxide. Partly because of this absorption, and partly because the temperature of the earth's surface is usually greater than that of the atmosphere, there is a net loss of energy from the earth's surface in the form of long wave radiation.

Figure 4 illustrates schematically the radiation balance at the earth's surface. In this diagram S_T is the total incoming solar radiation in both direct and diffuse forms, S_R is that solar radiation immediately reflected at the surface, while L_D and L_u are respectively the downward (or incoming) long wave radiation to the surface and the upward (or outgoing) longwave radiation from the surface. Solar or

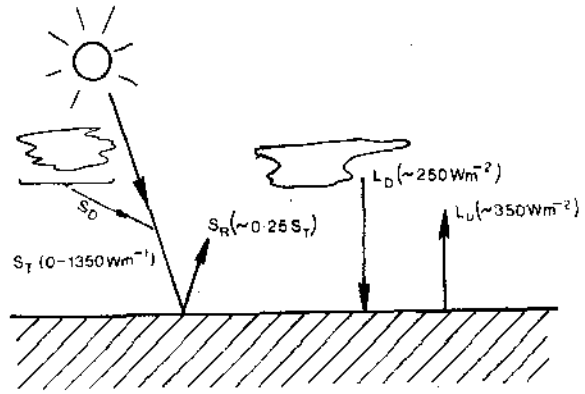

FIGURE 4

SCHEMATIC DIAGRAM OF THE RADIATION BALANCE AT THE EARTH's SURFACE. S_T IS THE TOTAL SHORTWAVE RADIATION, OF WHICH S_D IS THE DIFFUSE COMPONENT; S_R IS THE REFLECTED SHORTWAVE RADIATION, WHILE L_D AND L_u ARE THE INCOMING AND OUTGOING COMPONENTS OF LONGWAVE RADIATION

shortwave radiation is arbitrarily assigned the wavelength limits 0.3 to 3 μm, while longwave radiation is assigned the wavelength limits 3 to 30 μm. The amount of incoming solar radiation S_T, typically in the range 0 - 1350 W m^{-2}, varies with the altitude of the sun, and therefore varies with the time of day and time of year, as does the fraction of shortwave radiation reflected (although as already mentioned the daily average reflection coefficient or albedo is very commonly approximately 0.25 for short crops). The upward longwave radiation varies with surface temperature but is typically in the order of 350 W m^{-2}, while the downward longwave is typically in the order of 250 W m^{-2}. The difference, $L_u - L_D$, is usually in the range 87 to 107 W m^{-2} in clear sky conditions. The situation in cloudy conditions is more complex; for a more extensive description of radiation exchange in general, and longwave radiation exchange in particular, the reader is referred to pages 14 to 77 of Monteith (1972).

2. DIFFUSION

Basic Concept

Diffusion is the process by which those properties which can be used to characterize a particular volume of fluid, *e.g.* its heat content, momentum content or the concentration of its constituent parts, are moved from one position in the fluid to another. Such movement takes place when there are variations in the concentrations of these characteristic entities from one position to the next, and occurs because, at normal temperatures, the molecules making up the fluid are in permanent and random motion, either as individual molecules or as coherent groups of molecules. There is a random exchange of molecules or groups of molecules between adjacent positions and the characteristic entity is transferred along with this mass exchange. If the concentration of the entity at these two positions was originally different there will be a net transfer, or flow, in response to the different concentration.

Molecular Diffusion

Molecular motion in air is extremely rapid: molecules have a root mean square velocity of about 500 m s^{-1} but have a mean free path between collisions of around 5 x 10^{-8} m. Molecular diffusion occurs as a direct consequence of this molecular agitation: the rapid molecular movement is responsible for the transfer while the high collision rate ensures the rapid equilibration of individual molecules with mean air properties at each new location.

It is a plausible assumption, confirmed by experiment, that the rate of flow, or 'current', of an atmospheric entity is directly related to the gradient in the concentration of that entity. The transfers of most relevance to an understanding of the evaporation process are those of momentum, heat and water vapour. If we consider transfer in one dimension (parallel to an arbitrarily chosen z axis) and consider the transfer of water vapour, then the transfer expressed as a mass flux per unit area, E, is proportional to the concentration gradient (in kg m^{-3}) according to the equation

$$E = - D_v \frac{\partial x}{\partial z} \qquad (3)$$

in which D is a *molecular diffusion coefficient* with dimensions of m^2 s^{-1}.

In an analogous way the concentration of heat in air is related to air temperature, T, through the volumetric heat capacity (ρc_p), where ρ is the density of air and c_p its specific heat, so that the rate of heat transfer H per unit area is given by the expression

$$H = - D_h \frac{\partial (\rho c_p T)}{\partial z} \qquad (4)$$

Similarly, if u is the velocity of the air perpendicular to the z axis, then the air possesses momentum in this plane with a concentration (ρu) and there can be a momentum transfer, giving rise to a viscous force or shearing stress per unit area, τ, with

$$\tau = D_M \frac{\partial (\rho u)}{\partial z} \qquad (5)$$

Because the same process is responsible for all of these different transfers, the diffusion coefficients D_V, D_H and D_M are similar in size, around 0.18 $m^2\ s^{-1}$, and all increase by about 0.7% per ^{o}C at normal temperatures.

It is very common to treat the evaporation rate in its equivalent form as the flow of latent heat. In this case equation (3) is usually rewritten in terms of the latent heat flux, λE, which is related to the gradient of atmospheric vapour pressure by the equation

$$\lambda E = - \frac{\rho c_p}{\gamma} D_V \frac{\partial e}{\partial z} \qquad (6)$$

in which γ is the so called psychrometric constant, being the combination of terms ($c_p \rho / \lambda$ 0.622) and having a value of 0.66 mb $^{o}C^{-1}$ at 20 ^{o}C and 1000 mb. It occurs frequently in this work; its use here gives rise to simplification later in the analysis and enhances similarity with equations (4) and (5) which can be re-written as

$$H = - \rho c_p D_H \frac{\partial T}{\partial z} \qquad (7)$$

and

$$\tau = \rho D_M \frac{\partial u}{\partial z} \qquad (8)$$

to a good approximation.

3. TURBULENCE

Introduction

The wind blowing horizontally over a natural surface is retarded by an interaction with the surface. In principle very smooth surfaces do exist in nature but in practice virtually all natural surfaces are aerodynamically 'rough', even at low windspeeds. The interaction of moving air with a rough surface gives rise to an apparently random and haphazard movement in which portions of air of varying size are continuously being created and destroyed, but move in an ill-defined yet coherent way during their transient existence. This phenomenon,

known as turbulence, is initiated by the non-uniformity at the surface, but propagates upwards into the atmosphere in a way determined by the height above the ground. The mixing it generates is a very efficient mechanism for transferring entities through the atmosphere away from the surface, much more efficient than simple molecular diffusion. It is the primary process responsible for the exchange of mass, momentum and heat between air close to the ground and that at higher levels in the atmosphere.

Much of the turbulence is produced by the frictional retardation of the wind, but the transfer properties of this frictional turbulence are enhanced or diminished if there is a gradient in mean air temperature along a direction away from the surface. In the absence of any mean wind at all, turbulent transfer would still occur by free convection whenever temperature differences existed in the atmosphere. In general,frictional turbulence, generated by the interaction of the wind with the surface, and free convection exist together in a hybrid form known as 'mixed convection'. It is usual to treat this situation theoretically by treating frictional turbulence as the primary transfer mechanism with the effect of free convection as a semi-empirical correction. In recognition of this, it is convenient here to consider first the situation in a 'neutral' atmosphere, that is one in which there is no mean temperature gradient away from the surface.

(a) Forced convection

The windspeed u(z) measured at a height z above an extensive horizontal surface uniformly populated with roughness elements, usually vegetation, is found to increase linearly with ln (z-d) in neutral conditions. The behaviour is illustrated in Figure 5. The parameter d can be regarded as a characteristic of the surface cover, to be determined by experiment

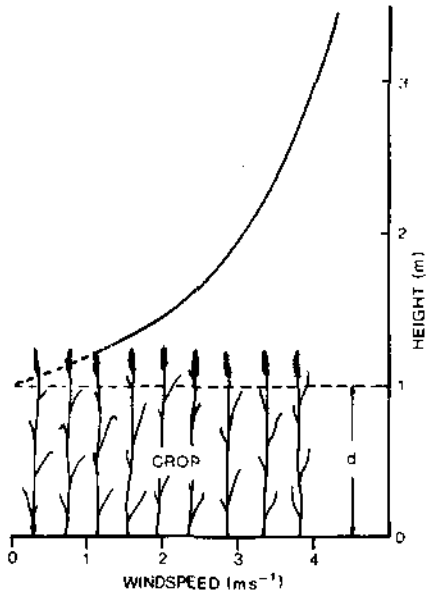

FIGURE 5

SCHEMATIC DIAGRAM OF THE VARIATION IN MEAN WINDSPEED OBSERVED OVER AN EXTENSIVE, UNIFORM STAND OF VEGETATION IN NEUTRAL CONDITIONS. THE PROFILE SHOWN FOLLOWS EQUATION (9) WITH d = 1 m AND z_o = 0.1 m

and known as the *zero plane displacement*. It is found to vary with both the height of the roughness elements, h, and their spacing, but is very commonly between 0.6 h and 0.8 h for agricultural crops and plantation forest. The empirical observation of a logarithmic wind profile can be expressed formally as

$$U(z) = k_3 \left[\ln(z-d) - \ln z_o \right] \tag{9}$$

in which z_o, like d, is characteristic of the surface and is called the *roughness length*. Both d and z_o can be found in practice by fitting equation (9) to experimental measurements of the windspeed profile measured in the field, with d, z_o and k_3 as free parameters. Such measurements must be made when there is no temperature gradient and this can be difficult: at least three accurate measurements of mean windspeed are required, and preferably several more. The roughness length of natural surfaces is usually in the order of 0.1 h.

In much the same way that the transfer of momentum by molecular diffusion is represented by equation (5), it is usual to describe the transfer by turbulent diffusion in the same way, with an equation of the form

$$\tau = K_M \frac{\partial u}{\partial z} \tag{10}$$

in which K_M is the *eddy diffusivity* for momentum, and is analogous to D_M, the molecular diffusion coefficient for momentum. The turbulent boundary layer through which this turbulent exchange takes place, is defined as the height range through which τ is constant. Equations (9) and (10) applied in this boundary layer, together require that K_M must be proportional to (z-d) *viz*

$$K_M = k_4 (z-d) \tag{11}$$

From equations (9), (10) and (11), it follows that τ must be given by the equation $\tau = k_3 k_4$ in which the product $(k_3 k_4)$ has dimensions of $(\text{velocity})^2$. It is usual to rewrite this product as u_*^2 with u_* called the *friction velocity*. In order that this equivalence should always be valid, it is necessary that both k_3 and k_4 must be proportional to u_* and that constants of proportionality should be reciprocals *viz* $k_3 = k\, u_*$ and $k_4 = u_*/k$. From the above argument it follows that

$$\tau = \rho\, u_*^2 \tag{12}$$

while equation (9) becomes

$$u(z) = \frac{u_*}{k} \ln \frac{z-d}{z_o} \tag{13}$$

The factor k in this equation is a scaling constant, a property of the turbulent wind field, which is assumed independent of the details of

the surface. It is known as *von Karman's constant* and is usually assigned the value 0.41.

It follows that the turbulent transfer coefficient K_M is given by

$$K_M = k\, u_* \,(z-d) \tag{14}$$

or from equation (13) as

$$K_M = k^2 (z-d) \; \partial u / \partial \ln (z-d) \tag{15}$$

which can be determined from the logarithmic gradient of the measured wind profile, once d is known.

In just the same way that K_M is defined from equation (10), as the eddy diffusivity for momentum transfer, it is possible to define similar entities for the transfer of sensible and latent heat from the equations

$$H = - \rho c_p K_H \frac{\partial T}{\partial z} \tag{16}$$

and $$\lambda E = - \frac{\rho c_p}{\gamma} K_V \frac{\partial e}{\partial z} \tag{17}$$

which have obvious analogy with equations (6) and (7). The turbulent transfer mechanism responsible for the exchange of all of these entities is considered the same. It is usual to assume $K_M = K_H = K_V$ are the same over the height range of the near-surface turbulent boundary layer *providing* there is no temperature gradient away from the aerodynamically rough surface.

(b) Mixed Convection

When there is a finite gradient of temperature away from the surface, the transfer by forced convection, described in the previous section, is modified. When air temperature decreases with height, any parcel of air created and moved upwards in the forced convection process will tend to continue its ascent, because it will be warmer, lighter and therefore more buoyant than the air into which it is moved. In this situation it is usual to describe the atmosphere as '*unstable*'. On the other hand, when the temperature of the atmosphere near the surface increases with height the opposite occurs. In this case buoyancy forces act against the further upward motion of a similar parcel of air, and the atmosphere is said to be '*stable*'.

The behaviour of the turbulent transfer mechanism in stable and unstable conditions differs from forced convection in neutral conditions, and it is to be expected that this difference might most reasonably be described empirically in terms of parameters which reflect the relative efficiency of the free and forced convection mechanisms. Free, or buoyancy generated, turbulent transfer is related to the *square* of the windspeed gradient. For this reason enhancement (or otherwise) of

forced convection in other than neutral conditions has very often been parameterised in terms of a dimensionless parameter R_i, known as the Richardson number, which is related to the ratio of these production mechanisms and is defined by

$$R_i = \frac{g}{T} \frac{\partial T}{\partial z} / \left(\frac{\partial u}{\partial z}\right)^2 \qquad (18)$$

where g is the acceleration due to gravity and T is the temperature in degrees Kelvin.

The transfer coefficients for momentum, and sensible and latent heat in a general mixed convection situation are redefined from equation (13) as

$$K_M = k\, u_* (z-d)\, \phi_M \qquad (19)$$

$$K_H = k\, u_* (z-d)\, \phi_H \qquad (20)$$

and $$K_V = k\, u_* (z-d)\, \phi_V \qquad (21)$$

The functions ϕ_M, ϕ_H and ϕ_V express the stability dependent description of mixed convection as a modification of forced convection, and are empirical. In stable conditions the best available data *e.g.* Webb, 1970; Mann, 1960; and Lumley and Panofsky, 1964) indicate an inter relationship of the form

$$\phi_V = \phi_H = \phi_M = (1 - 5R_i)^{-1} \quad \ldots\ R_i\text{+ve} \qquad (22)$$

while in unstable conditions, Dyer and Hicks (1970) concluded that

$$\phi_V = \phi_H = \phi_M^2 = (1 - 16R_i)^{-\frac{1}{2}} \quad \ldots\ R_i\text{-ve} \qquad (23)$$

These factors are included via equations (19), (20) and (21) as empirical corrections in the one dimensional diffusion equations (10), (16) and (17).

4. RESISTANCE

Although it is sometimes convenient to express the movement of energy (as latent and sensible heat) with the differential diffusion equations described in the previous section, it is a common and useful simplification to apply these equations in an integrated form. The integration is very straightforward in cases where the (one-dimensional) flux can be treated as a constant in the direction specified by the coordinate z. If equation (6) is used to describe the molecular diffusion of latent heat between two points at which the vapour pressure is e_1 and e_2 respectively, then this description can be integrated to the form

$$\lambda E = -\frac{\rho c_p}{\gamma} \frac{e_2 - e_1}{\int_{z_1}^{z_2} \frac{dz}{D_V}} \qquad (24)$$

Clearly the integration is independent of the detailed mechanism responsible for the diffusion and equation (17) can be treated in exactly the same way to give

$$\lambda E = -\frac{\rho c_p}{\gamma} \frac{e_2 - e_1}{\int_{z_1}^{z_2} \frac{dz}{K_V}} \qquad (25)$$

Equations (24) and (25) have an obvious similarity to Ohm's Law for electrical current *viz*

$$\text{Current} = \frac{\text{Potential Difference}}{\text{Resistance}}$$

and on the basis of this analogy it is usual to use the name 'resistance' for the entities $\left[\int_{z_1}^{z_2} \frac{dz}{D_V}\right]$ and $\left[\int_{z_1}^{z_2} \frac{dz}{K_V}\right]$ in these equations. Obviously similar resistances can be defined for the transfer of both momentum and sensible heat, with D_V replaced by D_M or D_H for molecular diffusion, and K_V replaced by K_M or K_H for turbulent diffusion.

The introduction of the concept of 'resistance' into models of the exchange between vegetation and the atmosphere allows a description which is mathematically similar at each stage in the transfer path, even though the physical mechanism changes from molecular diffusion, close to the surface, to turbulent diffusion, in the atmospheric boundary layer. It is convenient here to discuss each of the physical processes in turn and interpret the process in terms of an equivalent resistance.

(a) Atmospheric Resistance

In a previous section the turbulent transfer away from the aerodynamically rough earth surface was described in terms of the eddy diffusivities K_M, K_H and K_V. Clearly, on the basis of equation (25), the equivalent resistances between any two levels z_1 and z_2 follows directly from the equations

$$r_M^{1,2} = \int_{z_1}^{z_2} \frac{dz}{K_M(z)} \qquad (26)$$

$$r_H^{1,2} = \int_{z_1}^{z_2} \frac{dz}{K_H(z)} \qquad (27)$$

$$r_V^{1,2} = \int_{z_1}^{z_2} \frac{dz}{K_V(z)} \qquad (28)$$

When a multilevel description of the vegetation-atmosphere exchange is attempted, which includes the turbulent diffusion between different levels in or above the canopy, then this is the formulation of aerodynamic resistance which is used. However, a more common requirement is the bulk turbulent transfer resistance between some effective source level, in the vegetation, and z_h, the level in the atmosphere at which meteorological measurements of windspeed, temperature and humidity are available. This is usually taken to be the total turbulent transfer resistance between the reference level z_h, and $(z_o + d)$, the level at which the logarithmic wind profile, extrapolated into the canopy from above, would predict zero windspeed in neutral conditions (Figure 5). The value for the bulk transfer resistance for momentum in a neutral atmosphere is obtained by combining equations (13) and (14) and integrating between $(z_o + d)$ and z_h : it has the form

$$r_M = \frac{\left[\ln (z-d)/z_o \right]^2}{k^2 \, u(z)} \qquad (29)$$

It is usual to assume that the bulk transfer resistances for sensible and latent heat, r_H and r_V, are equal to r_M in neutral conditions, *i.e.* that

$$r_H = r_V = r_M \quad .... \; R_i \approx 0 \qquad (30)$$

If the atmosphere is not neutral it is necessary to take account of the stability corrections described in equations (18) to (23) and associated text. For a more complete description of the relationships between the bulk aerodynamic resistance for different properties, the reader is referred to Stewart and Thom (1973).

(b) Boundary-Layer Resistance

Fluxes of momentum and latent and sensible heat arise at the surface of

the individual elements making up the canopy. Moving air within the vegetation interacts with these elements and the flux is exchanged through a 'boundary-layer resistance' which, in the case of sensible and latent heat, might be envisaged as the result of molecular diffusion through a boundary layer close to the surface of the element.

The situation for momentum is rather different since in this case the transfer can occur at enhanced rates as a result of the additional interaction of pressure forces on the element. In consequence of this, the boundary layer resistance for momentum, r_{bM}, is usually significantly less than that for the heat fluxes, r_{bV} and r_{bH}. The effective boundary layer resistance for momentum is defined by the equation

$$r_{bM} = \rho \frac{\bar{u}}{\tau} = \frac{\bar{u}}{u_*^2} \qquad (31)$$

where $\bar{u}$ is the 'mean canopy windspeed' (Thom, 1972).

The additional aerodynamic resistance seen by the sensible and latent heat fluxes (which cannot be transferred by pressure interactions), the so called 'excess resistance', r_B, is conventionally expressed in terms of the non-dimensional parameter B^{-1} (Owen and Thompson, 1963; Chamberlain, 1966; 1968) such that

$$r_B = \frac{B^{-1}}{u_*} \qquad (32)$$

The size and variation of the excess resistance have been studied by Chamberlain (1966) and Thom (1972). These studies suggest that the value of B^{-1} (typically in the order of 4) is not strongly related to surface roughness but depends on u_*. Thom (1972) further suggests that an empirical relationship of the form

$$B^{-1} = C\, u_*^{1/3} \qquad (33)$$

'may well provide accurate enough first approximations', with C in the order of 5.6, when u_* is in units of m s^{-1}.

If the excess aerodynamic resistance is mainly in the boundary layer, then the boundary layer resistance for sensible and latent heat may be estimated from that computed for momentum using the relationship

$$r_{bH,V} = r_{bM} + r_B$$

$$= r_{bM} + C\, u_*^{-2/3} \qquad (34)$$

(c) Stomatal Resistance

The sensible heat and momentum fluxes can, with reason, be considered to

originate on the surface of vegetation elements making up the canopy. However if the canopy is dry, the latent heat flux arises in the first instance as a result of evaporation from cell walls inside stomatal cavities within the vegetation. It escapes to the leaf surface by molecular diffusion through the stomatal pores. Once at the surface the water vapour can diffuse through the leaf boundary layer and then into the atmosphere along the same path as momentum and sensible heat. The additional diffusion through the stomatal aperture represents an additional resistance, which is only applicable to latent heat, and is called stomatal resistance. It is not relevant in wet conditions when the vapour source is on the surface of the vegetation.

A typical leaf is usually considered small enough for temperature gradients across its surface to be ignored. The atmosphere in the stomatal cavities within the leaf can be assumed to be at saturation vapour pressure corresponding to leaf temperature. Water potentials in leaves are rarely below - 40 bars, and at this potential the equilibrium relative humidity is still 97% of saturation. With this simplification it immediately follows that the flow of water vapour to the leaf surface, *i.e.* the transpiration, can be described by the equation

$$\lambda E = \frac{\rho c_p}{\gamma} \frac{e_s (T_o) - e_o}{r_{ST}} \qquad (35)$$

where r_{ST} is the stomatal resistance for unit surface area, and T_o and e_o are the temperature and vapour pressure immediately adjacent to the leaf surface.

5. NATURAL EVAPORATION

(a) The Energy Budget

The size of the total heat flux ($H + \lambda E$) from a plant community is strictly limited to the energy available from the current input to the community, *i.e.* to the 'available energy'. This is defined to be the energy input to the vegetation as radiation minus any energy removed in directions other than vertically upwards, or stored somehow within the community.

In general the region over which incoming energy is partitioned extends over a finite height range between the soil surface and the level at or above the vegetation at which radiation is measured. Within this region temperature changes occur in both the atmosphere and the vegetation, and changes occur in the absolute humidity of the air. Such changes represent a loss or gain in the energy available for partition into latent and sensible heat for any particular period, the magnitude being proportional to the rates of change of temperature and humidity with time. Energy lost or gained in this way is called 'storage', S; it is usually quite a small proportion of the total energy budget except at dawn and dusk for tall vegetation, and is often very close to zero when integrated over the normal diurnal temperature cycle.

Within the region of interaction, energy is absorbed or released by the biochemical processes involved in photosynthesis and respiration within the plants. Such 'biochemical storage', P, is usually quite small and has been estimated as about 2 per cent of net radiation in daytime conditions (Stewart and Thom, 1973).

A component of the energy usually of more importance is the energy lost by thermal conduction into the ground. This energy flux, the 'soil heat flux', G, represents a loss of energy during the day and again at night time. It is very significant at particular times of day for short vegetation and bare soil, if the diurnal cycle in surface temperature is large, and can be as high as 150 W m^{-2}: it is often negligible for forest stands (Sinclair *et al* 1975). The integrated value over a day is normally fairly small, but the long term integrated value can still have some influence on seasonal evaporation rates.

Although the majority of the energy as sensible and latent heat leaves normal to the earth's surface, horizontal gradients of temperature and humidity observed across the ground necessarily imply that some of this energy is leaving parallel to the surface as advection. This horizontal flux divergence, D, is generally neglected in energy balance studies, often with justification, but short range advection can be an important term in 'oasis' situations, and the effect of medium and large scale advection is under-rated, particularly for forest vegetation (Shuttleworth and Calder, 1979).

Collecting these several terms together, the available energy, A, for a complete stand of vegetation is given by

$$A = R_N - D - G - S - P \qquad (36)$$

where R_N is the net radiation input to the stand per unit area.

In mathematical models of the vegetation/atmosphere interaction involving energy partition at several different levels in the canopy, it is necessary to consider the energy available at each height. The radiation intercepted at each level normally forms the basis of this calculation, with the additional energy terms outlined above included where relevant. For a more complete treatment of this topic the reader is referred to Sinclair *et al* (1971) and Shuttleworth (1976).

(b) Resistance Networks

(i) Multilayer Models

The network of resistances involved in the partition of available energy into sensible heat and evaporation is complex for real stands of vegetation, but successful computer models have been built which describe the process in one dimension (e.g. Waggoner and Reifsnyder, 1968; Sinclair *et al*, 1971). Such models involve the simultaneous, iterative solution of energy balance equations at several heights in the canopy, and require estimates of the resistances operating at each level, *i.e.* the stomatal

and boundary layer resistances, as well as the effective 'eddy' diffusion resistances operating between levels. Figure 6 illustrates the network of resistances commonly used.

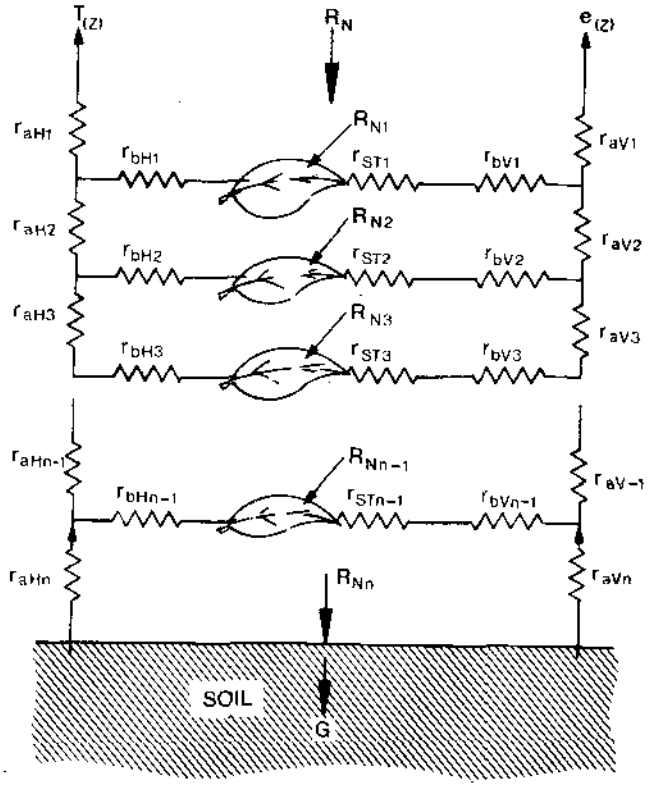

FIGURE 6

DIAGRAM OF THE RESISTANCE NETWORK USED TO PARTITION ENERGY INTO LATENT AND SENSIBLE HEAT IN 'MULTI-LAYER' MODELS OF THE VEGETATION/ATMOSPHERE INTERACTION. THE RESISTANCE r_{STn} IS THE STOMATAL RESISTANCE APPLICABLE AT EACH HEIGHT, WHILE r_{bHn} AND r_{bVn} ARE THE BOUNDARY LAYER RESISTANCES. r_{aHn} AND r_{aVn} ARE THE EDDY DIFFUSION RESISTANCES BETWEEN LEVELS

Typically the model starts by calculating the available energy relevant to each height, and then uses the assumed resistance chain to calculate canopy profiles of temperature and humidity. The procedure is described in some detail by Waggoner (1975). Given correct input data, such models have been used to provide a realistic description of detailed behaviour by simulating measured behaviour in real canopies. The virtue of this work is that it provides reassurance that the basic physical elements, described in the previous sections of this report, can indeed be successfully combined to describe the observed behaviour of real canopies.

Use of such models is unlikely to advance the basic physical understanding used in their synthesis, but they do provide a check on whether this understanding is sufficiently advanced. The input data required restricts their predictive use; however, once tested, such models can provide extremely useful yardsticks against which to test simpler models more suitable to predictive application *(e.g.* Sinclair *et al;* 1976). This point is considered in section III 2(a).

(ii) Single Source Models

In practice it is generally possible to provide a useful and realistic physical description of the evaporation process without needing to resort to the use of a multilevel description. It is often found that it is possible to combine the in-canopy resistances, the stomatal and boundary layer resistances, assuming they acted in parallel at a single level in the canopy. The resulting bulk stomatal and boundary layer resistance, r_{ST} and r_B respectively, are then considered to act at an 'effective source height' somewhere close to the apparent sink of momentum. In single source models of this type it is no longer relevant to separate the 'boundary layer' and 'eddy diffusive' resistances, since they are now assumed to act in series on all the fluxes involved; it is more usual to combine them as a single 'aerodynamic' resistance, r_A, which is the bulk transfer resistance mentioned at the end of section 1 4 (a).

Figure 7 illustrates a single source model for the partition of energy into latent and sensible heat fluxes. The latent heat flux still differs from the sensible heat flux in that it is subject to the additional stomatal resistance, r_{ST}, in dry canopy conditions. The temperature

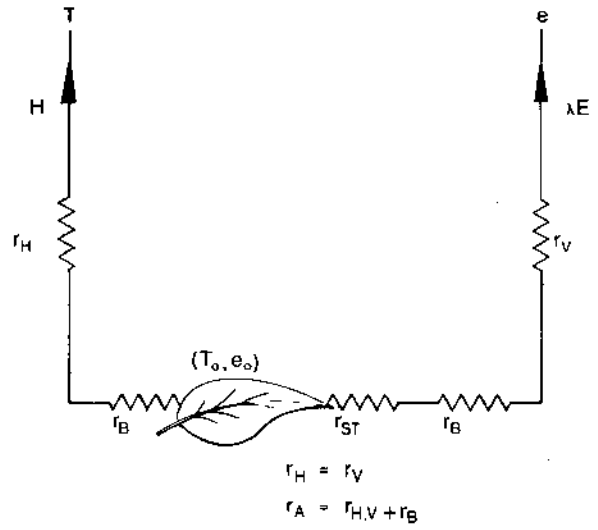

FIGURE 7 SCHEMATIC DIAGRAM OF THE RESISTANCE NETWORK USED IN 'SINGLE-SOURCE' MODELS OF THE VEGETATION/ATMOSPHERIC INTERACTION. r_{ST} AND r_B ARE THE BULK STOMATAL AND BOUNDARY LAYER RESISTANCES; r_H AND r_V ARE USUALLY TAKEN AS EQUAL AND COMBINED IN SERIES WITH THE BOUNDARY LAYER RESISTANCE TO GIVE THE 'AERODYNAMIC' RESISTANCE r_A ($r_A = r'_H + r_B = r'_V + r_B$). r_A IS THEN THE 'BULK TRANSFER RESISTANCE', GIVEN BY EQUATIONS (29) AND (30)

and humidity are measured at a 'screen height' above the vegetation, and have the values T and e respectively ; the effective temperature and humidity adjacent to the dry leaf surface are T_o and e_o. Inside the stomata the air is saturated with a vapour pressure $e_s(T_o)$. According to this model the latent heat flux is given by

$$\lambda E = \frac{\rho c_p}{\gamma} \frac{(e_s(T_o) - e)}{r_A + r_{ST}} \tag{37}$$

while the sensible heat is given by

$$H = \rho c_p \frac{(T_o - T)}{r_A} \tag{38}$$

Using the mean gradient of the saturated vapour pressure curve defined in equation (2), it follows from equation (38) that

$$H = \frac{\rho c_p}{\Delta} \frac{(e_s(T_o) - e_s(T))}{r_A} \tag{39}$$

while energy conservation requires that

$$H = A - \lambda E \tag{40}$$

Substituting this last equation into equation (39), and eliminating $e_s(T_o)$ between the resulting equation and equation (37) gives the result

$$\lambda E = \frac{\Delta A + \frac{\rho c_p}{r_A}(e_s(T) - e)}{\Delta + \gamma(1 + \frac{r_s}{r_A})} \tag{41}$$

This equation, generally called the Penman-Monteith equation, is the fundamental expression used in simple one-dimensional descriptions of the evaporation process, and is the basis of all the more empirical techniques used in estimating evaporation. This is discussed in more detail in the context of evaporation estimation in section III of this report.

II EVAPORATION MEASUREMENT METHODS

As we have already mentioned, the process of natural evaporation is the change of phase from liquid water to water vapour which occurs over a limited height range near the interface between the atmosphere and the earth's surface in response to the energy input from the sun. Its measurement can therefore either be made in the liquid phase, as the rate of loss of liquid water from the surface, or in the gaseous phase, as the rate of gain of water vapour by the atmosphere. In general, measurements in the liquid phase assume or create a 'closed' system beneath but near to the earth's surface, and deduce evaporation as the net loss of water from that closed system over a given time, the measurement being fundamentally one of *discrete changes* in total quantity. On the other hand measurements in the vapour phase assume that the atmosphere away from the surface represents an 'open' system, and determine evaporation as an integration of the *rate of flow* of water vapour (or equivalently latent heat) into that open system through the turbulent boundary layer near the surface.

1. MEASUREMENTS OF LIQUID WATER LOSS

All measurements of this type rely on drawing up a mass (volume) balance for the water content of a specified volume of soil beneath the earth's surface. The surface area of this sample is a necessary part of the measurement while its depth can either be well defined, as in lysimetric measurements, or poorly defined but large enough for vertical drainage to be neglected, as in catchment experiments. This difference influences the timescale over which the results have worthwhile accuracy. In each case the measurement reduces to determining the terms in a basic water balance equation of general form

$$E = P - (V_R + V_S + V_L)/A \qquad (42)$$

where E is the total evapotranspiration loss from the specified volume per unit area

P is the precipitation (irrigation) input to the specified volume per unit area

V_R is the total volume of liquid water leaving the specified volume as measured 'runoff' both above and below the surface

V_S is the change in the total volume of liquid water stored within the specified volume

V_L is the 'leakage', *i.e.* that total volume of liquid water leaving the specified volume which is not, or cannot be, measured; and which therefore represents an error in the method.

A is the effective area of the sample volume at the earth's surface.

The terms in this equation are discrete quantities of water determined over a common time interval.

All water budget measurements share the problem that the error in the evapotranspiration calculated from equation (42) is an accumulation of the errors in the other measured variables, and a worthwhile result requires that these variables should be known with fairly high accuracy. This problem is compounded by the fact that E is often only a fraction of P, so that its calculation from equation (42) is commonly as a small difference between several large terms. Water balance methods include catchment hydrology, soil moisture depletion sampling and lysimeters, all of which we outline separately.

(a) Catchment Hydrology

The design of catchment experiments capable of yielding a worthwhile measurement of evaporation is a subject too broad to allow adequate description in a report of this nature, and already forms a significant part of hydrological literature. The interested reader is referred to such literature for more detailed information, with the book 'Systematic Hydrology' by Rodda *et al* (1976) recommended as a useful, recent account. In the context of this work it is sufficient only to draw attention to some of the more important difficulties involved.

The largest uncertainty in the evaporation loss deduced from a catchment water balance is the possibility that the unmeasured leakage, V_L, forms a significant part of the total water movement. Considerable skill is required in selecting natural catchments without leakage: it is important to recognise that the subterranean groundwater contours play an important, perhaps definitive, role in specifying catchment boundaries, and that surface topography is not necessarily a reliable reflection of subsurface flow.

It is also important to remember that the storage term V_S in equation (42) is very often difficult to measure reliably in extensive natural catchments and will usually provide the most important error in any attempt at a short term evaporation measurement. Its significance becomes less for a long term determination of average evaporation, when the error in this component can become comparable with those in precipitation and runoff.

In the light of the very real possibility of significant error in the bulk evaporation loss deduced from a catchment experiment, it is

recommended that any such measurements are substantiated by parallel and independent meteorological or lysimetric measurements.

(b) Soil Moisture Depletion

Given adequate and reliable instrumentation capable of providing measurements of soil moisture content in sufficient number to reduce sampling errors, and providing there is little natural drainage from the near surface soil volume over which such instrumentation is used, then the determination of V_S in equation (42) can provide an indirect measurement of evaporation when precipitation is infrequent and light. A neutron probe, such as that described by Bell (1969, 1973), is the most practical means of providing the necessary determination of changes in moisture content, because it gives a direct measurement of water volume fraction for soil samples of reasonable large volume which can be measured repeatedly and *in situ*.

The method can only give worthwhile measurements over periods longer than a week and even over this timescale percentage changes in total water content are usually small. The possibility that significant drainage might occur severely limits the technique, unless steps are taken to limit measurements to a volume of soil from which drainage is known to be slight. The simultaneous measurement of soil water tension profiles with tensiometers such as those described by Cooper (1978) can significantly enhance the usefulness of the method, by defining the vertical position of a point of inflection in the tension gradient through which vertical water movement should not occur. The integrated measurement of moisture change above this level can then provide a more realistic measurement of the total evaporation loss between sample periods. However, the method is still only applicable in locations and/or over periods where subterranean water movement occurs by the process of unsaturated flow.

(c) Lysimetry

A lysimeter is a device in which a volume of soil, which may be planted with vegetation, is isolated hydrologically from the surrounding soil. They are constructed to make $V_L = 0$, and either to permit measurement of V_R, or to make it zero. If the lysimeter is to provide useful measurements of E the sample volume of soil and its vegetation cover must be representative of the surrounding area. Steps must be taken to ensure that the thermal, hydrological and mechanical properties of the soil are similar, and to establish that the vegetation sample is representative in terms of height, density and physiological well-being. A well designed and maintained lysimeter can be of considerable use, not only in providing measurements of E, but also as an independent check on micrometeorological methods and for the calibration of empirical or semi-empirical formulae. Unfortunately the cost of installing satisfactory lysimeters, and the level of maintenance they require, generally restricts their use to research applications with specific or local relevance.

In setting up a lysimeter it is best if the soil sample is obtained by casing a block of soil, at least as deep as the rooting depth, *in situ;* although in coarse sandy soils it is sometimes possible to 'refill' the lysimeter with systematic attention to the original profile of soil type and packing density. If the lysimeter is fairly shallow it is preferable to instal 'suction control' at the bottom of the soil sample so as to maintain a soil moisture tension equal to that at the same level in the adjacent, free-draining soil. In general the surface area of a lysimeter should be large compared to the scale of the surface and subsurface inhomogeneities and care should be taken to minimize the differences in the thermal properties inside and outside the lysimeter. The thermal conductivity of the surface layer is particularly important for short term measurements, and the temperature near the base of the sample important for long term measurements. If the sides of the lysimeter are exposed to the atmosphere they should be sealed to stop evaporation from the soil sample, while, in weighing lysimeters, care should be taken to avoid condensation on the containing walls by drying the air surrounding the tank.

The most reliable weighing lysimeters use mechanical balances to measure the water loss from a soil/vegetation sample contained in a tank, which is itself inside a second tank, to retain the surrounding soil. Success has however been obtained with floating lysimeters and hydraulic load-cell lysimeters; the reader is referred to Gangopadhyaya *et al* (1966) for a useful resumé of weighing techniques.

In nonweighing lysimeters the evaporation from a sealed sample of soil ($V_L = 0$) is deduced from equation (42) as the difference between precipitation and runoff ($P - V_R/A$) over medium or long time scales (when the V_S term is small); or as $(P - (V_R + V_S)/A)$ over a shorter (weekly) timescale, if the stored water compoment can be measured with a neutron probe. In certain restricted conditions the presence of an impervious layer in the soil profile favours the construction of a nonweighing (or 'natural') lysimeter of this type by removing the need to seal the base of the soil sample (e.g. Calder 1976).

(d) Plant Physiological Techniques

In addition to the three more commonly used techniques, there exists a fourth class of measurements, which fall naturally into the category of liquid water loss measurements, but differ in the respect that they measure only transpiration. They provide a measurement of actual evaporation only when this is the primary process involved in the total evaporation loss; nevertheless they have a significant role to play, particularly in those situations where water use is of most importance *viz* in arid climates, in irrigation control and in drought conditions.

Such methods involve measuring the rate of water uptake from a representative sample of the vegetation. Successful measurements have been made of transpiration in field conditions by cutting the stems of the vegetation under water, immersing the cut end in a water tank and noting the water uptake (*e.g.* Roberts 1977). Such experiments require frequent checks on the plant physiological parameters involved in transpiration control, particularly stomatal resistance, to ensure that the sample remains

representative of nearby vegetation, and they can clearly only be used over periods which are short compared to the growth cycle of the vegetation involved. Measurements of the rate of water flow up the stems of representative vegetation which do not require disturbing the flow by cutting the stem are preferable. This usually involves introducing 'tracers' into the stem of the vegetation to give a measure of the rate of flow of sap through the conducting tissue : tests of this method using radioactive phosphorus as a tracer have recently proved successful (Waring and Roberts; 1979).

2. MEASUREMENTS OF WATER VAPOUR TRANSFER

The ability of the atmosphere to transfer water vapour in both vertical and horizontal directions is so great that it is not practical to measure evaporation by determining the change in vapour content of a closed sample of air of finite extent. For this reason meteorological, or more correctly micrometeorological, measurements of evaporation differ fundamentally from those using liquid water loss, in that they determine evaporation from a consideration of the dynamical processes responsible for vapour transfer away from the surface: the measurement is of the rate of flow, not of the net transfer.

Measurements are made in the atmosphere, within the turbulent boundary layer close to the surface, so that the measured upward vapour transfer rate is a very good approximation to the surface exchange rate. Much closer to the surface vapour transfer occurs by molecular diffusion, but within the turbulent boundary layer the turbulent exchange process is much more efficient and is the dominant exchange mechanism. There are two broad classes of micrometeorological evaporation measurement, the diffusion equation approach and the eddy fluctuation approach. Both rely, in one respect or another, on the fact that one dimensional turbulent transport is the primary transfer mechanism over a height range above, but not too far from, the surface.

Natural evaporation depends on the type and form of the surface and for this reason alone there can be significant horizontal variation in its magnitude. Since micrometeorological measurements are necessarily some distance above the ground, and the atmosphere is almost always moving with respect to the earth, the measurements obtained at a particular location are representative of an area some distance up-wind. For an eddy correlation measurement this can be an advantage in that the turbulent mixing in the atmosphere up-wind of the instruments helps to produce a result which is the average evaporation over a fairly large area: the only restriction is that the surface inhomogeneities should not be large enough to produce persistent flow patterns where the mean wind is no longer parallel to the evaporating surface. However, if the evaporation is meant to be representative of a particular uniform crop type, or if one of the diffusion equations is to be used, it is necessary that there should be an extensive 'fetch' of evaporating surface with essentially identical properties extending upwind from the measurement site to a distance at least 50-100 times the height of the micro-meteorological instrumentation.

(a) Diffusion Equation Measurements

The basic assumption behind the measurement of evaporation by diffusion equation techniques is that above an extensive homogeneous surface the transfer of atmospheric entities, such as water vapour, can be described by a vertical, one-dimensional diffusion equation. It is further assumed that the turbulent diffusion coefficients relating the fluxes of water vapour, sensible heat and momentum to the respective vertical gradients of humidity, temperature and windspeed are the same; or that they are at least related to each other in a way which is not determined by characteristics of the evaporating surface, rather by characteristics of the turbulent boundary itself, and that this relationship is therefore more likely to be universal.

Strictly speaking, describing the vertical transport in this simple way, also implicitly assumes that the diffusion equation represents a steadystate, dynamical exchange, in which the fluxes and gradients are constant. In fact this is very rarely the case, but the timescale of the change in these entities is usually significantly longer than the timescales involved in the turbulent transfer mechanism itself. In practice, the average gradients over periods of 20 to 60 minutes can be successfully used to deduce the average fluxes over this time period.

(i) *Aerodynamic Methods*

Over an aerodynamically rough, flat surface the horizontal windspeed, u, has a vertical dependence of the form

$$u(z) = \left[(\tau/\rho)^{\frac{1}{2}}/k \right] \{ \ln(z - d)/z_o + \Phi \} \qquad (43)$$

where τ is the momentum flux

ρ is the density of air

k is a constant ($\simeq$ 0.41), called 'von Karman's' constant

d and z_o were parameters which can be determined from the windspeed profile and are called the 'displacement height' and '*aerodynamic roughness parameter*' *respectively*

and Φ is a function of the Richardson number, R_i, which is itself given by equation (18).

The fluxes of momentum (τ), sensible heat (H) and latent heat (λE) are assumed to be related to vertical gradients of windspeed (u), potential temperature (T) and vapour pressure (e) by one dimensional diffusion equations of the form

$$\tau = -\rho K_H \frac{\partial u}{\partial z} \; ; \quad H = \rho c_p K_H \frac{\partial T}{\partial z} \; ; \quad \lambda E = \frac{\rho c_p}{\gamma} K_V \frac{\partial e}{\partial z}$$

where K_M, K_H and K_V are the eddy diffusivities. The ratio of any two of

these fluxes is proportional to the ratio of the difference in their related gradients, so that when measurements are made at two levels z_1 and z_2, the fluxes of sensible heat and latent heat can be deduced from their ratio to the calculable momentum, using equations which take the final form

$$H = \frac{-\rho c_p k^2 (K_H/K_M)(T_2 - T_1)(u_2 - u_1)}{\ln^2[(z_2 - d)/(z_1 - d)]} \frac{1}{\phi^2} \qquad (44)$$

$$\text{and } \lambda E = \frac{-(\rho \frac{c_p}{\gamma}) k^2 (K_V/K_M)(e_2 - e_1)(u_2 - u_1)}{\ln^2[(z_2 - d)/(z_1 - d)]} \frac{1}{\phi^2} \qquad (45)$$

where ϕ is a 'universal' stability correction (related to Φ by $d\Phi/d \ln (z - z_o) = \phi - 1$) which on the basis of present experimental evidence might be assigned the form given in equations (22) and (23).

Given the assumption $(K_H/K_M) = 1$ and $(K_V/K_M) = 1$, it is possible to apply equations (44) and (45) to deduce evaporation, providing measurements are made of d, $(u_2 - u_1)$, $(e_2 - e_1)$ and $(T_2 - T_1)$. To find d it is necessary to make measurements of windspeed at several (at least three) levels in conditions where there is no atmospheric temperature gradient (*i.e.* $\Phi = 0$), and to determine d (and z_o) by optimizing equation (43). In practice very high accuracy is required because the determination of d depends on the second order properties of the profile. Its value is commonly about three quarters of the height of the vegetation cover; its precise value is less important in equations (44) and (45) if z_1 and z_2 are large with respect to the vegetation height, but this is often difficult to achieve because of fetch requirements.

This technique is limited by the assumption that K_H and K_V are equal to K_M, which means that it should not be used in very unstable conditions ($R_i >> 0$) over short vegetation, and it should not be used at all near the surface of tall vegetation when these assumptions are not valid (Thom *et al*, 1975).

In addition to the basic aerodynamic method just outlined, there are several adaptations which, in general, attempt to reduce or alter the experimental requirements in an attempt to simplify the implementation of a practical measurement. One of the most important of these is the *Deacon and Swinbank method* in which three anemometers are used, two well above the crop surface and the third near the surface, where the windspeed profile is not strongly influenced by thermal stratification. If the two high anemometers are far enough from the surface for d to have a negligible effect in equation (45), then these two measurements alone are sufficient to determine τ from equation (43) in conditions when there is no temperature gradient ($\Phi = 0$). Such measurements in turn allow the determination of a bulk drag coefficient, C_s, for the near surface windspeed, u_s, using the equation

$$\tau = C_s \rho u_s^2 \qquad (46)$$

If it is assumed that this drag coefficient is also valid in other conditions, since the windspeed measurement is close to the surface where thermal effects are small, then it is possible to calculate τ in all conditions and deduce a value of the evaporation flux from the ratio of the differential measurements of windspeed and vapour pressure made at the higher levels *viz*

$$\lambda E = - C_s \left(\frac{\rho c_p}{\gamma}\right) u_s^2 (e_2 - e_1)/(u_2 - u_1) \qquad (47)$$

(ii) *Energy Balance Methods*

Whereas aerodynamic methods deduce vapour flux from its ratio with respect to momentum flux, energy balance methods deduce it from its ratio with respect to sensible heat flux. The technique uses measurements of temperature and humidity at coincident levels to determine the ratio of the sensible and latent heat fluxes, the Bowen ratio β, from the equation

$$\beta = \frac{H}{\lambda E} = \gamma \frac{\Delta T}{\Delta e} \qquad (48)$$

where γ is the so-called psychrometric 'constant' $\left[c_p \, p/(\frac{8}{5}\lambda) \approx 0.66 \text{ mb}^{\circ}\text{C}^{-1} \right]$
The ratio of the vertical gradients of temperature and humidity can either be determined, as in the above equation, in the form of a differential measurement (e.g. McNeil and Shuttleworth, 1975), or as the ratio of the tangents to fitted temperature and humidity profiles (e.g. Stewart and Thom, 1973).

Having determined the ratio of the two energy fluxes in this way, it is then necessary to measure the sum of these fluxes, the 'available energy' A, *viz*

$$A = H + \lambda E \qquad (49)$$

in order to calculate the evaporation from equations (48) and (49); which can be combined in the form

$$\lambda E = A/(1 + \beta) \qquad (50)$$

To obtain A it is necessary to draw up an energy budget for a volume of unit cross-section which includes all the evaporating surfaces participating in the exchange. The assumption of horizontal homogeneity simplifies this by restricting attention only to vertical fluxes and changes in total energy content. The volume over which the energy budget is relevant is that between the level at which the incoming net radiation is measured, and the level, just beneath the soil, at which the soil heat flux is measured. Within this volume the contributions to the total energy budget can be summarized as:

R_N - the net input of radiation of all wavelengths, measured with a net-radiometer;

G - the heat flux into or out of the ground, measured with soil heat flux plates inserted just below the soil surface;

S_A - the net change in energy stored as sensible or latent heat in the volume of air under consideration, calculated from the rate of change of the measured temperature and humidity;

S_B - the net change in energy content of the vegetation if present, estimated from sample measurements of biomass, and assumed or measured vegetation temperature;

P - the net energy absorbed or released by the combined processes of photosynthesis and respiration within the plant community, always small ($\lesssim$ 2 per cent of net radiation) and often negligible.

The available energy is the linear sum of the average value of these several terms over the averaging time used for the meteorological measurements, *viz*

$$A = H + \lambda E = R_N - G - S_A - S_B - P \qquad (51)$$

The energy budget measurement technique is superior to the aerodynamic method in that the background assumptions are less and is preferred for this reason. There are periods where $A \approx 0$ and $\beta = -1$ when the method fails, and evaporation estimates are required during these low flux conditions.

(iii) *Combination Methods*

Both the aerodynamic and energy budget methods as described above require measurements of the humidity content of the air at different levels in order to deduce the evaporation flux, and in practice this is probably the most difficult measurement required. Techniques have been used which interchange sensors between different levels on a short time scale in order to reduce systematic errors in the measurements (e.g. McNeil and Shuttleworth, 1975). Amongst the several types of measurements described in the category of Diffusion Equation Measurements, the energy budget measurement with interchanged sensors is probably the preferred method. If this is not feasible for technical reasons, then it is possible to make a measurement of evaporation by a combination of the aerodynamic and energy budget methods which avoids the need for humidity measurements. The technique is to determine H by the aerodynamic method using equation (44), and then to deduce evaporation from the energy budget using equation (51). This combination method is still limited by the technical difficulties involved in making an accurate determination of d, and still requires the theoretical assumptions necessary for the aerodynamic method.

(b) Eddy Correlation Measurements

Within the turbulent boundary layer generated near the ground by the interaction of wind with surface irregularities, the transfer of water vapour and other atmospheric entities occurs by the process of turbulent diffusion. Although the mean wind at each level near the surface is of course parallel to the ground, the instantaneous wind at any particular point can assume any direction and, in general, there is a component of wind intermittently away from or towards the surface with a mean value of zero, but a finite standard deviation. It is this fluctuating component of wind which transports the atmospheric entity away from or towards the surface as a suspended feature of the moving air. In order to get a net transport there must also be fluctuations in the concentration of the suspended entity, and these fluctuations must be correlated in some degree with those in the windspeed. In this way an outward evaporation flux occurs when, on average, air movement away from the surface is correlated with higher than average humidity, and air movement towards the surface is correlated with lower than average humidity. Evaporation can be measured by making simultaneous measurements of both the fluctuations in the wind component normal to the surface and fluctuations in the humidity content of the air at the same point. The algebraic product of the magnitude of these fluctuations is the instantaneous vapour flux, which can be either positive or negative at any instant (in accordance with the statistical nature of the turbulent transfer process), but which has a finite mean value, the net evaporation loss from the surface, found by integrating the instantaneous flux.

If the time dependent component of windspeed normal to the surface is represented by $w(t)$ and the specific humidity of the air by $q(t)$ then the mass of water vapour passing through unit area in an element of time dt is given by

$$E(t) = \rho\, w(t)\, q(t)\, dt \qquad (52)$$

where ρ is the density of air. The average quantity of moisture transferred through this plane parallel to the evaporating surface in unit time, the evaporation, will be $\overline{\rho w q}$ (*i.e.* the mean value of $\overline{\rho w q}$). Denoting the mean value of these components by a bar and fluctuating components by a prime, the evaporation is then given by

$$\begin{aligned} E &= \rho w q = \overline{(\rho w) q} \\ &= \overline{((\overline{\rho w}) + (\rho w)')\ (\bar{q} + q')} \\ &= \overline{\rho w}\ \bar{q} + \overline{\overline{\rho w}\, q'} + \overline{(\rho w)' \bar{q}} + \overline{(\rho w)' q'} \end{aligned}$$

by definition the second and third terms in this equation vanish leaving

$$E = \overline{\rho w}\ \bar{q} + \overline{\rho w' q'} \qquad (53)$$

The term $\overline{\rho w}\ \overline{q}$ represents the mean transfer through the reference level an is usually assumed to be zero, unless there is some evidence to suggest that the mean air flow is not parallel to the evaporating surface. The term $\overline{\rho w'q'}$ is the eddy flux of water vapour which is to be measured with fast response sensors to yield a direct measurement of evaporation.

It is clearly possible to repeat the above analysis with specific humidity replaced by temperature to yield a similar expression for the sensible heat flux *viz*

$$H = \overline{\rho c_p w}\ \overline{T} + \overline{\rho c_p w'T'} \qquad (54)$$

where c_p is the specific heat of air. This equation can also be used as the basis of a measurement of evaporation in combination with a specification of the energy budget using equation (51).

The practical problems involved in using the eddy correlation method routinely are two fold:

(i) To create sensors capable of making measurements of the fluctuating windspeed and humidity (or temperature) at rates high enough to include all the higher frequencies participating in the turbulent transfer while at the same time being stable enough to give a reliable measurement of the lower frequency components of this transfer. The actual frequency range required varies with the size of the vegetation above which the system is to operate, its height above the ground and the ambient windspeed, since these affect the scale of the turbulence in the turbulent boundary. In practice a sensor capable of operating between 10 and 0.001 Hz should be capable of producing the necessary measurements in most real applications.

(ii) To carry out a fairly complex analysis on the sensor outputs in real time with data sampled at a time interval which is half the sensor time response. This problem has been greatly reduced with the advent of large scale integrated circuits and the consequent availability of cheap digital processors.

Successful measurements of atmospheric fluxes have been made using the eddy correlation principle but these have usually formed part of more extensive, fundamental studies of near surface atmospheric turbulence (e.g. Kaimal *et al*, 1972). Such work has necessitated 'state of the art' sensors and data acquisition systems, requiring a level of technical competence to operate and maintain the system and also generating stored data in quantities which are not consistent with the single simple objective of evaporation measurement *per se*. Attempts have been made to develop simple eddy correlation systems (*e.g.* Hicks, 1970), usually employing a single propeller anemometer as the wind sensor, and these have been used, although with less than total success (*e.g.* McNeil and Shuttleworth, 1975; Moore, 1976).

In its basic simplicity the eddy correlation measurement of evaporation is the most elegant of the meteorological methods; it is the method with the minimum of theoretical assumptions and with the least dependence on surface conditions. Its large scale application is limited only by the technical difficulties in applying it in a routine and reliable way. Ultimately the method should prove the most accurate and straightforward means of measuring evaporation but at the present time the instrumental problems involved in its application are not solved.

Here at the Institute, we have embarked on a project to re-evaluate the eddy correlation method as a routine measurement of evaporation in light of the technological progress that has occurred in the last few years. On the basis of a literature survey (Moore, McNeil and Shuttleworth, 1976), initial attention is being centred on cheap, sonic anemometry and propeller anemometry (with on-line correction) as prospective sensors of wind vector; and on a simplified infra-red absorption device (with on-line correction for sensor drift) as a prospective sensor of humidity fluctuations. Evaluation of platinum wire thermometers, thermocouples and thermistors as potential sensors of temperature fluctuations, for use in an eddy correlation/energy budget combination method, is also in progress. The intention is to use a microprocessor-based logger/processor system to carry out the necessary real-time calculations in the field.

III EVAPORATION ESTIMATION METHODS

1. INTRODUCTORY CONCEPTS

The multiplicity of techniques purporting to estimate evaporation is daunting, and the variation in their type and form disconcerting to those unfamiliar with the problem. This complexity is compounded by the fact that 'estimation equations' are commonly used in two modes; firstly in the prediction of likely evaporation on the basis of an assumed correlation with existing data; and secondly as a means of avoiding the measurement of actual evaporation, by replacing it with simpler, but hopefully related, measurements. Several estimation equations can be used in either of these modes. The situation is further complicated by the fact that workers in this field have found it useful to create 'standard' evaporation rates, namely *Potential Evaporation*, *Potential Evapotranspiration* and *Reference Crop Evaporation*, which attempt to provide some measure of the atmosphere's ability to support the evaporation process. Some of the estimation techniques attempt to estimate one or more of these conceptual entities rather than actual evaporation. We make some attempt at clarification by expanding briefly on these points.

Predictive Estimation

The range of applications in which evaporation estimates are required in general exceeds the data from which they can be made with any definitive and unique physical basis. This situation tends to force the use of correlation equations, generated on the basis of previous measurements, relating evaporation to some (usually meteorological) parameters available in the original study and also available in the present application. Such equations are always statistically correct expressions of a real, albeit indirect, correlation, but sometimes have limited physical or physiological basis and therefore dubious universality. Nevertheless the procedure is perhaps justifiable, even from a scientific standpoint, since it does represent an approximate solution to a practical problem which probably has at least some background justification, even if this is improperly understood. The reliability and universality of a predictive estimation equation increases as it becomes a closer approximation to a physical description of the actual evaporation process; unfortunately the data requirement necessary for the estimate also increases at the same time.

Estimation 'Measurements'

The need for an observer to generate a measurement based number, which he can present as a determination of evaporation, often surpasses the technological ability and financial resources available to supply it. Faced with this difficulty, experimentalists have very often preferred

to make other simpler measurements and attempted to use these in empirical or semi-empirical relationships with evaporation. This is difficult to justify from a strict scientific standpoint, although justification can usually be made on the grounds of finance or expediency. The creation and perpetuation of a large number of empirical or semi-empirical evaporation 'measurement' techniques is a direct consequence of the fact that, historically, the measurement of natural evaporation has proved difficult and expensive; indeed this situation still persists despite current technological progress. Given this unfortunate state of affairs, it is hard to criticize the use of 'estimation measurement' methods although constant vigilance is necessary to ensure that the philosophy does not prove counter productive by diverting resources and attention away from the real problem, that of evaporation measurement, into the side problem of measuring input parameters to empirical equations.

Standard Evaporation Rates

It is clear from the physical description of the evaporation process given earlier in this report, that the rate of natural evaporation can be a complex function of atmospheric, soil and vegetation factors. In an attempt to try to bring some structure into this complexity, it has generally been considered convenient to attempt to define certain 'standard' rates of evaporation, which are designed to give a measure of the meteorological or climatological control on the evaporation process at a particular location. These standard rates are conceptual, in the sense that they represent idealised situations; they also maintain a level of imprecision in their definition in those aspects, particularly related to the surface, where this is considered advantageous.

Penman (1948) created the concept of *Potential Evaporation* which might be defined as, 'the quantity of water evaporated from an idealized, extensive free water surface per unit area, per unit time under existing atmospheric conditions'. Although this definition is imprecise, in that it does not specify certain criteria which would influence such evaporation in a real situation (*eg* the extent of energy transfer within the water or the level of turbulent transfer above the water surface), it is a conceptual entity which can be easily visualized, which will bear some reasonably close relationship to actual evaporation from particular free water sources, and which will give some measure of the meteorological control on evaporation.

On the basis of the experimental evidence available, it was for many years believed that the type and form of vegetation cover on the earth's surface (even whether it was there or not) had little effect on the rate of natural evaporation, *providing* this was limited by the heat supplied to the surface and not by the availability of surface water. In presence of this belief it is reasonable to conceive an entity, *Potential Evapotranspiration*, which might be conservative at a particular location and determined mainly by meteorological conditions. It can be defined (Gangopadhyaya *et al*; 1966) as, 'the maximum quantity of water capable of being lost, as water vapour, in a given climate, by a continuous, extensive stretch of vegetation covering the whole ground when the soil is kept saturated'. It includes both evaporation from the soil and from the vegetation for a specified region over a given time interval. The concept was, and is, used as a

'scale' upon which the influence of surface control can be superimposed, usually as a multiplication factor, often parameterized in terms of soil moisture deficit. The term remains in common use today, and still has some validity as a conservative entity serving as a climatological term upon which medium term measurements of actual evaporation in different geographical locations can be related. It is increasingly recognised however, that, in order to preserve the entity's essential conservative nature as a local meteorological scale, it is necessary to acknowledge an effect of vegetation type and even local position, to the extent that any empirical estimation formulae used to describe it require local calibration where possible (Tanner, 1967).

Gradually it has become clear that the initial observation of a lack of dependence on vegetation cover when water is non limiting, and the idea that potential evapotranspiration represents a maximum rate (which implies that energy advection is a scarce and transient phenomenon), might be related to the fact that many of the original studies took place over short crops. With such crops the control exerted by the atmosphere itself is maximised, since it dictates not only the driving potential in the diffusion process, but also generates the dominant, possibly controlling, resistance to vapour transfer. This realization, coupled with the continuing desire to create a conservative but surface independent entity, has given rise to the recreation of a better defined standard evaporation rate, *Reference Crop Evapotranspiration* (Doorenbos and Pruitt, 1977) (*Note.* It should be remembered that in Penman's *original* work, λE_T referred to the evaporation from *turf*). This entity is defined as 'the rate of evapotranspiration from an extensive surface of 8 to 15 cm tall, green grass cover of uniform height, actively growing, completely shading the ground and not short of water'.

There is a considerable overlap between the concept of Potential Evapotranspiration and Reference Crop Evapotranspiration, particularly in regard to the empirical formulae used to estimate them, but the better definition of Reference Crop Evaporation avoids the problems of vegetation control and advection, and therefore increases the probable universality of locally derived empirical equations.

All these 'standard' evaporation rates are meant to be a measure of meteorological control at a particular location with as little dependence on surface effects as possible: they choose to ignore the fact that meteorological parameters near the surface are themselves influenced by up-wind surface exchange rates. In this respect it should be remembered that if they are to fulfil their intended rôle and, in the form of a medium term average, remain conservative entities at a given location, there is an implicit assumption that there will be no drastic change in average surface cover for some distance up wind. In the same vein, it can be noticed that the definition of these standard rates refer to 'extensive' rather than 'infinite' surfaces; this is in recognition of the fact that in the infinite limit, never achieved in practice, the surface exchange controls the behaviour of the atmosphere and gives rise to an evaporation rate

$$\lambda E = \frac{\Delta}{\Delta + \gamma} (R_N - G) \qquad (55)$$

called 'equilibrium evaporation'. This represents the lower limit of evaporation from any 'wet' surface, a fact recognised by Slatyer and McIlroy (1961) and recently formalised by McNaughton (1976).

2. THE CLASSIFICATION RATIONALE OF ESTIMATION TECHNIQUES

The extensive range of estimation techniques presently available makes it difficult at first sight to discern any order or hierarchy in their form and input requirements, and confuses their physical relationship where this exists. A recent and useful piece of work by Stewart (1979) has helped to classify a broad group of techniques by making explicit their interrelationship through the meteorological input they contain. Here we draw on and extend this classification in an attempt to show that it is possible to include most of the methods based on individual meteorological parameters. There remains however an anomalous subgroup not depending on meteorological information, in which a direct measurement is made of the rate of evaporation from a free water surface *e.g.* evaporation pans. Measurements of this last type are extensively used, usually in an 'estimation measurement' mode, as a means of estimating potential evaporation (in the first instance). They do not fall sequentially into a hierarchy based on individual meteorological inputs: they are best considered as a measurement of the complex combination of meteorological inputs directly relevant to a particular and fairly simple evaporating system.

Part of the purpose of attempting a classification scheme is to attempt a partial explanation of why some of the formulae, with little apparent physical basis, might give rise to a valid statistical correlation with empirical constants of at least local significance. The classification, and the hierarchy it imposes, is also useful as a basis for understanding why particular estimation techniques can, in the first instance, only realistically provide direct estimates of one of the standard evaporation rates rather than actual evaporation. It is a common procedure to generate guesses of actual evaporation by first estimating one of these conceptual standard rates, and then introducing a second level of empiricism in the form of a multiplicative constant, to create a number which can be presented as an estimate or even 'measurement' of actual evaporation. There is very little physical basis for this procedure, although it is easy to understand conceptually and popular: as yet it has not been openly reported as giving rise to catastrophic practical consequences in real applications. We outline the procedure later in the text.

The classification scheme outlined in Table 1, attempts to itemize the various evaporation estimates together with their data requirement, the type of evaporation they might legitimately provide, *eg* Actual Evaporation, Potential Evaporation *etc*, and a subjective assessment of the level of current usage and whether this is growing or declining. The scheme is meant to be fairly general in terms of the broad classes described, but no attempt is made to include every example of the numerous alternative forms within each class. This is partly because

doing so would tend to confuse rather than clarify, and partly because it is the author's opinion that argument regarding the relative merits of individual examples in each class is very often semantic in light of their empirical basis.

The reader familiar with evaporation estimation will recognise a definite and deliberate bias in this presentation towards newer techniques, not already comprehensively described in the literature in any unified way, whose use is sometimes speculative, and the subject of active development. In general the trend in current research is towards the development of techniques capable of estimating actual evaporation directly. This usually means that the method has to attempt a closer description of reality by explicitly acknowledging the effect of the surface in the technique itself. Such methods generally require a greater data input. The present usage of the various techniques is partly conditioned by familiarity and popular acceptability, and largely by the current availability of relevant historical data. It is observation that the popular acceptance of a particular estimation technique has often preceded the widespread availability of the data required for its use.

We now describe the various techniques in the descending order of Table 1, attempting to describe the additional assumptions involved in progressing from one class to the next as we do so.

(a) Simulation Models

Clearly the technique most likely to estimate actual evaporation is to build a model which as nearly as possible simulates the physical and physiological processes which actually occur in the real situation. This can only really be attempted with the availability of a large digital computer and in the presence of a high level of detailed understanding of the basic properties involved in the processes giving rise to the energy fluxes. Such models are perhaps best used as a test of the validity of this understanding rather than as a practical means of estimating bulk evaporation rates.

Usually models are built in one dimension and attempt to simulate evaporation from vegetation by including all available information regarding the vegetation stand under study, *eg* its structure and form, and submodels of its stomatal behaviour in response to meteorological parameters. The models can be supplied with short term measurements of the meteorological conditions as an input, and solve the equations relating the meteorological and physiological parameters simultaneously at all the levels in the canopy using these meteorological inputs as a boundary condition. In doing this they generate simulated profiles of meteorological parameters such as temperature, vapour pressure and the heat fluxes, which can then be compared with actual measurements as an indirect test of the understanding used in building the model and the validity of the submodels employed.

TABLE 1 (a)

	MODEL CLASS	EVAPORATION ESTIMATED
III B(a)	Simulation	
	(i) Numerical (SPAM)	Actual
	(ii) Analytic (Shuttleworth)	
III B(b)	Single source	Actual
	(i) Transpiration (Penman-Monteith)	
	(ii) Interception (Rutter)	
	(iii) Unified (Shuttleworth)	
III B(c)	Intermediate (Thom-Oliver-Gash)	λE_T
III B(d)	Energy Balance	λE_{RC}
	(*eg* Penman)	λE_T (short crop)
III B(e)	Radiation	λE_{RC}
	(*eg* Priestley-Taylor)	λE_T (short crop)
	[(Shuttleworth/Calder)]	[λE_T]
III B(f)	Humidity	λE_{RC}
	(*eg* Dalton)	λE_T (short crop)
III B(g)	Temperature	λE_{RC}
	(*eg* Blaney-Criddle)	λE_T (short crop)
III B(h)	'Direct' Methods	λE_o
	(i) Evaporation pans	(estimation 'measurement usage only)
	(ii) Atmometers	

TABLE 1(b)

	DATA REQUIREMENT	CURRENT USAGE
III B(a)		
	(i) Detailed models of physiological response (ii) Detailed information on canopy exchange processes (iii) Detailed information on canopy structure (iv) Short term measurements of meteorological data	Small, increasing interest
III B(b)		
	(i) Submodels of surface resistance (ii) Coarse measurements of canopy structure (iii) Short term measurements of meteorological data	Small, increasing usage
III B(c)	(i) Daily meteorological data (ii) Coarse measurements of canopy structure (iii) Information on rainfall pattern	Minimal increasing interest
III B(d)	Daily meteorological data (T, R_N, u, D)	Large, stable
III B(e)	Daily meteorological data (T, R_N) [Daily meteorological data (T, R_N,P)]	Medium scale, increasing [Minimal]
III B(f)	Daily meteorological data (T, u, e)	Small, decreasing
III B(g)	Daily meteorological data (T)	Medium scale, stable
III B(h)	Daily water loss measurements	Large, Stable

(i) *Numerical*

Several models exist which carry out a numerical simulation of a one-dimensional description of the vegetation/atmosphere, of the general form illustrated in Figure 6, by dividing the vegetation into a finite number of horizontal layers. About ten layers are usually used and for each layer the interception of solar and thermal radiation is calculated, and partitioned into sensible heat, latent heat and photochemical energy. Iterative procedures are used until an energy balance is achieved for all foliage layers. The SPAM model described in more detail by Sinclair *et al* (1971, 1976) is an excellent example. Such models have successfully described measured profiles and bulk energy fluxes and must be considered the best available method of predicting actual evaporation given an extremely high data availability, and *providing* the required submodels are available. It is not certain at present that universal, physiological submodels can be created, although considerable research effort has been and is currently being applied to the problem: it could be that fairly coarse submodels might suffice.

(ii) *Analytical*

Shuttleworth (1976) demonstrated that numerical multilayer models of the type just described can in fact be rewritten in analytical form by taking the limit corresponding to an infinite number of levels. He demonstrated that the use of such models are equivalent to using a formula of the form

$$\lambda E = \frac{\Delta A' + (\rho c_p D + \delta)/r_H}{\Delta + \gamma_E (1 + r_{c/r_V})} \qquad (56)$$

The definition of the terms in this equation, for which see Shuttleworth (1976, 1978), involves the integration of microclimatic variables through the canopy. It could be that there are some computational advantages in evaluating the bulk evaporation from this formula, using efficient integration techniques to evaluate the various integrals rather than using a multilayer model: this point remains untested. However iteration would still be required, and the technique would still be subject to the same basic shortcomings, namely the scarcity of short term data, the present non-availability of universal, physiological submodels, and the need for detailed measurements (or submodels) of crop structure and its variation in time.

The major objective in creating this analytical model was to demonstrate that the technique of numerical, multilayer simulation on the one hand, and the description of bulk evaporation by a single equation on the other, are actually, and not just numerically, equivalent. The major advantage is that it is possible to use the analytic nature of the model in a simplifying process to demonstrate the assumptions involved in using 'single source' models as a more practical alternative (Shuttleworth, 1978).

(b) Single Source Models

Single layer or 'big leaf' models of plant canopies (*eg* Monteith, 1965; Thom, 1972), assume that the overall effect of the whole canopy on the above canopy energy fluxes is reasonably approximated by a model which assumes that all the component elements making up the vegetation are exposed to the same microclimate. The sensible heat and latent heat (evaporation) fluxes flow through resistances r_H and r_V respectively in moving from the surface of the leaves at this level to some higher level above the canopy, 'the screen height', at which measurements of temperature and vapour pressure are made. In general the resistances r_H and r_V can differ from the equivalent resistance to momentum flux (Thom, 1972), and they can differ from each other, but in practice the assumption $r_H = r_V$ is a common and satisfactory assumption.

In dry conditions the latent heat flux, which arises inside the stomatal cavities when transpiration is taking place, is subject to the additional diffusive resistance, r_{ST}, it encounters in negotiating the stomatal opening. This resistance is usually called the bulk stomatal resistance. If the canopy is totally wet both fluxes arise at the surface of the vegetation, and the effective resistance to vapour exchange at a free water surface is negligible (Shuttleworth, 1975, 1978). The situation is more complex when the canopy is only partially wet and the simple extension of previous ideas has proved inadequate (Shuttleworth, 1975). However recent work (Shuttleworth, 1978) has suggested that an effective 'surface' resistance can in fact be created to describe the situation. So far this formulation has only been tested for tall (forest) vegetation (Shuttleworth, 1978). Should this success be repeated for short vegetation, then a unified single source model would exist for all rainfall conditions which might prove the basis for fairly reliable estimates of actual evaporation given the availability of short term (hourly) meteorological data and a realistic submodel of bulk stomatal resistance.

We discuss the totally dry canopy, totally wet canopy and universal descriptions separately.

(i) *Transpiration*

The single source representation of transpiring canopies is a well established and well tested model: it is already extensively described in the literature, see for example, Thom (1972). The hypothetical numerical similarity (Monteith, 1965) of bulk stomatal resistance and a vertical integration of the component stomatal resistances making up the canopy, has been tested experimentally, by Black *et al* (1970), Brun *et al* (1972), Szeicz *et al* (1973) and Tan and Black (1976); numerically, against a multilayer model, by Sincalir *et al* (1976); and theoretically by Shuttleworth (1976). The predicted evaporation given by this model takes the form

$$\lambda E = \frac{\Delta A' + \rho c_p D / r_H}{\Delta + \gamma \frac{(r_V + r_{ST})}{r_H}} \qquad (57)$$

where A' is the energy available for evaporation; often approximated by $(R_N - G)$, or even R_N, over long time periods and D is the vapour pressure deficit measured at 'screen height'.

In applying equation (57) it is very common to assume that r_H and r_V are equal and are related to r_M the bulk aerodynamic exchange resistance for momentum. The assumption

$$r_H = r_V = r_M = \frac{u(z)}{u_*^2} \qquad (58)$$

has been used with considerable success and will usually suffice in predictive applications within the likely accuracy of the other terms in equation (57), particularly r_{ST}. Thom (1972) argues that the resistances r_H and r_V should be calculated from

$$r_H = r_V = r_M + 6.3\, u_*^{-2/3} \qquad (59)$$

Care should be taken in using this last equation for tall vegetation unless some account is also taken of the, presently unexplained, anomalies in near surface turbulence above such vegetation (Thom *et al*, 1975). The effect of these anomalies is to compensate for the second term in equation (59), so that in practice equation (58) remains an adequate first order approximation.

The friction velocity, u_*, appearing in the above equations might be estimated in practical applications by

$$u_* = \frac{u(z)}{k} \ln \frac{z - d}{z_o} \qquad (60)$$

In practice d and z_o are quite commonly found to have an approximate relationship to h, the height of the vegetation, *viz* d = 0.75 h and z_o = 0.1 h, although care should be taken with smooth, dense crops (*eg* tea).

Given all the above assumptions it is possible to use equation (57) to estimate transpiration given an adequate submodel of r_{ST}, the bulk stomatal resistance. A great deal of research is currently investigating whether deterministic models can be made of stomatal resistance in terms of environmental variables both at the single leaf and whole canopy level (*eg* Jarvis; 1976). The effect of stomatal resistance is most pronounced for tall vegetation and several studies have been made of its behaviour. Stewart and Gash (1976) for example, found large daily variations but a consistent daily trend for a Scots Pine forest, while Calder (1977) was able to model the effective surface resistance of a Spruce forest by the expression

$$r_{ST} = 74.5 \left\{ 1 - 0.3 \cos \left(2\pi \left[\frac{N_D - 222}{365} \right] \right) \right\} / (1 - 0.45D) \qquad (61)$$

where N_D is the 'day number' in the northern hemisphere year. The equation is valid for values of D less than 2.2.

Clearly detailed plant physiological research such as that described by Jarvis (1976) should continue in an attempt to determine whether deterministic and universal submodels of bulk stomatal resistance do exist. However, it is the author's personal belief that such detailed short term modelling of this physiological entity in terms of a limited number of micrometeorological entities will not ultimately provide a universal relationship, even for an individual species, because it implicitly treats the stomata as purely physical systems and neglects their physiological and biochemical nature. At the same time it is also the author's belief that detailed short term submodels are not required for the particular problem of practical evaporation estimation, and that the problem is better solved by the accumulation and longer term modelling of large quantities of less detailed actual evaporation data from numerous species - although this would first necessitate the creation of simple techniques for measuring actual evaporation.

Some justification for this opinion can be obtained from equation (57). For tall vegetation, when $r_{ST} >> r_{H,V}$, the equation simplifies to the approximation

$$\lambda E = \frac{\rho c_p}{\gamma} \frac{D}{r_{ST}} + \lambda E_o \qquad (62)$$

'Pseudorandom' short term variation of r_{ST} with any other meteorological variable except D will tend to cancel out in a practical evaporation estimate; only variation with respect to vapour pressure deficit, and variation in response to longer term parameters, *eg* soil moisture deficit and the yearly cycle of plant growth, will have a large effect. At the same time, in the opposite limit of short vegetation and agricultural crops, the meteorological dependence of r_{ST} of primary importance is that with A' (R_N) and D. In this case the situation is further simplified by the fact that the effect of stomatal resistance is only as one of three terms in the denominator, all of similar magnitude. Setting $r_H = r_V = r_A$ in equation (57) and assuming r_{ST} (which can be slowly varying in response to longer term effects) has a fairly typical average value of 2 r_A, then a rather extreme change in r_{ST} to r_A and $3r_A$ will change the denominator of equation (57) from 2.7 to 2.1 and 3.4 respectively (for an assumed temperature of 20°C).

(ii) *Interception*

When the source of water vapour is a completely wet canopy of the vegetation, the term r_{ST} in equation (57) is zero. For the reasons

just outlined in the last section, the effect of a change in this parameter on the observed evaporation is most marked for vegetation types in which $r_{V,H}$ is small in comparison with r_{ST}; that is for tall (forest) vegetation; although the effect is still present to some degree for other vegetation types. Most work in modelling the evaporation of intercepted water has been concerned with forest stands where the necessary merging of such models with transpiration estimates is most simple (because here the evaporation rate for intercepted water is sufficiently high to ensure that partially wet canopies are of short duration).

The most successful model of forest interception is that due to Rutter *et al* (1971, 1975) which has been successfully tested by, for example, Gash and Morton (1978) and Calder (1977). The conceptual framework of the model is shown diagrammatically in Figure 8. In essence the model calculates a running water balance for the canopy and trunks of the forest stand using inputs of hourly rainfall and the meteorological parameters necessary to estimate evaporation. It computes the rate of evaporation of intercepted water, and also the amount of water reaching the ground in the form of drips from the canopy, 'throughfall', and down the trunks of the trees, 'stemflow'. The value of E_p in this diagram is calculated from equation (57) using the assumptions $r_H = r_V = U/U_*^2$ and $r_{ST} = 0$.

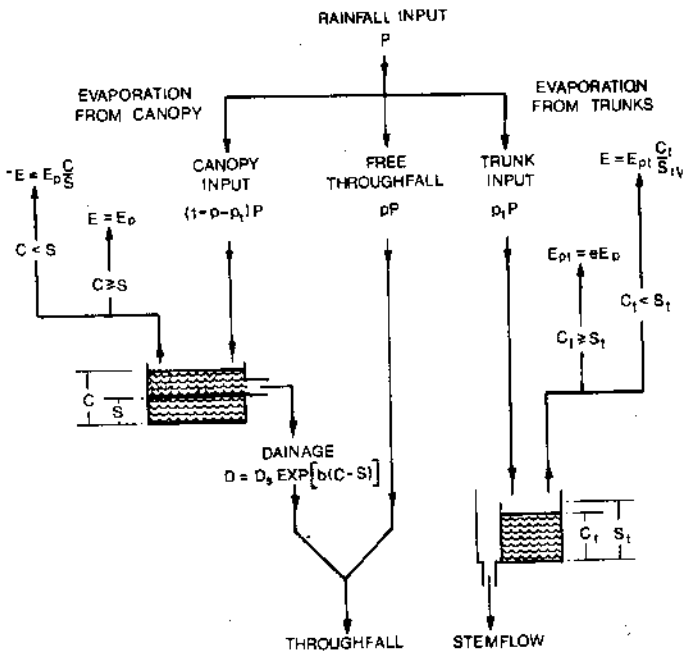

FIGURE 8 THE CONCEPTUAL FRAMEWORK OF THE RUTTER MODEL OF RAINFALL INTERCEPTION FROM FOREST VEGETATION DESCRIBED IN THE TEXT

The model uses C and C_t, the effective depth of water currently standing on the canopy and trunks respectively, as 'state variables' and requires a knowledge of the canopy structure and drainage in the form of the following parameters

S the canopy capacity: the amount of water left on the canopy when rainfall and throughfall have ceased (typically 1 mm).

p the free throughfall coefficient: the proportion of the rain which falls to the ground without striking the canopy (typically 0.25)

S_t the trunk water capacity (typically 0.1 mm)

p_t the proportion of rain diverted to stemflow (variable with species between 0.02 and 0.2, but with limited influence on the total evaporation loss)

D_S the drainage rate from the canopy, as throughfall, when the canopy storage is at capacity (typically 0.002 mm/min)

b an empirical coefficient used to calculate the throughfall drainage rate (typically 3.7 mm^{-1})

The drainage rate is given by the algorithm

$$\left.\begin{aligned} D &= D \exp\left(b\left[C - S\right]\right) \qquad C \geq S \\ &= 0 \qquad\qquad\qquad\qquad C \geq S \end{aligned}\right\} \tag{63}$$

The empiricism in the drainage calculation only influences the calculation of the evaporation through its effect on the amount of time taken to dry out the canopy after a rain storm: the drainage rate of a supersaturated canopy (dictated by the constant b) is commonly high, and the fall to canopy capacity quick, so that the time to drying out is mainly a function of the calculated evaporation rate (Gash, 1979).

For successful implementation of the model in an *estimation measurement* mode, short term (hourly) measurements must be made of rainfall and the meteorological parameters used in equation (57). If true estimation of the evaporation of interception loss is required, then these measurements must be available in the form of existing data sets. In order to estimate total forest evaporation a model of transpiration is necessary in parallel with this model. This must be 'turned off' while C retains a finite value since transpiration is suppressed while the canopy is wet (Gash and Stewart, 1976).

(iii) *Unified Single Source*

The overlap between the evaporation of intercepted water and transpiration is poorly defined in the model of forest evaporation just described. In order to unify the two processes correctly a physically continuous description of the transition between wet and dry canopies is required. In practice the less precise treatment outlined in the previous section probably provides an adequate estimate of total evaporation for forest stands, where partially wet canopies are of short duration. If single source estimation models of total actual evaporation are to be extended to short crops, then a realistic description of partially wet canopies is required. In any case a unified treatment is desirable for all crops providing this can be implemented in a simple and efficient way.

It is clear that equation (57) will form the basis of the unified single source model with r_{ST} replaced by a redefined surface resistance r_S which is equal to r_{ST} in dry conditions and equal to zero in wet conditions, and having a smooth transition between the two depending on some measure of the fractional surface wetness. A parameter similar to $(\frac{C}{S})$ in the Rutter model would obviously be the most convenient parameter for describing surface wetness. A recent theoretical analysis (Shuttleworth, 1978), so far only tested over tall vegetation, suggests the use of an expressing of the form

$$r_S = \left[\frac{W}{r_b\,(\frac{\Delta}{\gamma} + 1)} + \frac{W}{r_{ST} + r_b\,(\frac{\Delta}{\gamma} + 1)} \right]^{-1} - r_b\,(\frac{\Delta}{\gamma} + 1) \qquad (64)$$

$$\text{where } W = 0 \qquad C \geq S$$

$$= \frac{(R - 1)}{R - (\frac{C}{S}) - 1} \qquad C < S \qquad (65)$$

$$\text{in which } R = \left(\frac{r_{ST}}{r_A}\right)(r_A - r_b) \left[r_{ST} + (\frac{\Delta}{\gamma} + 1) r_b \right] \qquad (66)$$

with $r_A = r_H = r_V$ estimated as in the Rutter model and r_b an estimate of the mean *boundary layer* resistance of the vegetation elements in the canopy.

In this way, the unified single source model consists of a Rutter model, using C as a state variable to compute first W and then r_S, from equation (64), with a submodel of r_{ST}. The surface resistance r_S, which is a smoothly varying function of time, replaces r_{ST} in equation (57) to yield the required estimate of total evaporation.

(c) 'Intermediate' Models

Although simulation and single source methods are superior to all other techniques in that they provide a direct estimate of actual evaporation, their use, particularly in a *predictive estimation* mode, is limited by the current lack of short term meteorological data sets, and the submodels of stomatal resistance required for their implementation. An alternative way of estimating actual evaporation is by first estimating one of the standard evaporation rates and then attempting to modify these by an additional factor in an attempt to represent the real effect of surface vegetation. The remaining estimation techniques described here are primarily useful as a means of providing the initial estimate of such standard rates.

The Penman equation (Penman 1948), described in the next section, was initially created to provide an estimate of *Potential Evaporation*, λE_o, but Rijtema (1965) has demonstrated that, providing the measured or estimated net radiation used in the equation is that relevant to the

vegetation itself, it actually provides an estimate of *Potential Evapotranspiration*, λE_T, for *short green vegetation*. In the framework of the definitions used in this report, the Penman formula therefore provides an estimate of *Reference Crop Evapotranspiration*, λE_{RC}. The recognition of a distinction between λE_{RC} and λE_T is based on the realization that λE_T can differ from crop to crop in responce to different transpiration control and different levels of advective enhancement. All models previously described as estimating λE_T are more correctly described in the context of this work as providing estimates of λE_{RC}.

An estimate of the more variable Potential Evapotranspiration would however provide a better basis for the estimation of actual evaporation, and Thom and Oliver (1977) have attempted to modify the original Penman formula to provide an equation including additional terms with a physical basis, which could be adjusted empirically to yield an estimate of λE_T for different crops. The equation they propose takes the form

$$\lambda E = \frac{\Delta Q_N + \gamma E'_a}{\Delta + \gamma(1+n)} \qquad (67)$$

where E'_a is a modified version of the equivalent term in the Penman equation, namely

$$E'_a = 13.8\ (e_s - e)(1 + U/_{100})/\ln^2(z/z_o) \qquad (68)$$

where e and e_s (mb) are the actual and saturation values of vapour pressure measured at a height z above d, the zero plane displacement of the crop (d ~ 0.75 h where h is the vegetation height); U is the corresponding wind run, in miles per day. z_o is an estimate of the aerodynamic roughness parameter for the vegetation (of order h/10). Q_N, which appears in this equation and recurs in the next two sections is an estimate of the net energy available for evaporation at the evaporating surface. Q_N is most commonly a representation of $(R_N - G)$ in this work, where G is variously interpreted as the heat flux conducted downwards into the soil or free water surface.

The term n in equation (67) is a measure of the control exerted from the crop by virtue of its stomatal resistance. It is defined as $n = r_s/r_a$ in which r_a is the effective aerodynamic resistance, already included in E'_a, and given by the expression

$$r_a = 4.72\ \left[\ \ln^2\ (z/z_o)\ \right](1 + 0.54\ u) \qquad (69)$$

and r_s is the effective surface resistance estimated by

$$r_s = \left(1 - \frac{I}{E}\right) r_{sd} \qquad (70)$$

In this expression r_{sd} is an estimate of the average stomatal resistance for the crop, which is assumed constant. I is that part of the total evaporation, E, occurring directly from intercepted precipitation.

It is in this last equation that the Thom and Oliver approach loses some of its predictive potential by virtue of the fact that the parameter r_s, correctly, is not purely related to stomatal resistance. Obviously equation (67) can be calibrated against existing measurements of actual evaporation, to yield a value of r_{sd} and a time dependent function (I/E): r_{sd} might be relevant in another uncalibrated application, but it is not clear that (I/E) will necessarily be so.

Gash (1978) pointed out that the Thom and Oliver function could be rewritten into a form where it could be used in conjunction with separate estimates or measurements of interception loss, to provide a worthwhile predictive equation. Presently such interception estimates are only really available for tall vegetation. The equation he proposes has the form

$$\lambda E = \frac{\Delta Q_N + \gamma E'_a}{\Delta + \gamma(1 + r_{sd}/r_a} + I(1 - c) \qquad (71)$$

where in the additional term, I is the interception loss, and c is a correction term which compensates for calculating the transpiration even under wet conditions, and is given by

$$c = (\Delta + \gamma)/\left[\Delta + \gamma\ (1 + r_{sd}/r_a\ \right] \qquad (72)$$

Clearly equation (71) is only relevant if the estimate of interception does require short-term meteorological data, otherwise the method is merely another example of the single source technique using a constant value for the surface resistance. Recently Gash (1979) has proposed the use of a simple model of interception for tall crops which, neglecting evaporation from the trunks, takes the form

$$I = P_s + nS + \left(\frac{\bar{E}}{\bar{R}}\right)(P - P_s) \qquad (73)$$

where P is the precipitation input

P_s is the fraction of precipitation in rain storms less than $S/(1-p)$

S is the canopy storage

n is the number of storms with precipitation greater than $S/(1-p)$

$\bar{R}$ is the mean rainfall rate in storm conditions

and $\bar{E}$ is the mean evaporation rate from a totally wet forest canopy in storm conditions at the location under study

The three terms in this equation correspond to small storms (totally lost

because they do not wet the canopy), to evaporation after storms and evaporation during storms respectively. Such a description obviously requires some knowledge of the rainfall pattern for the area and estimates of the parameters p and S used in the Rutter model. It implies that $\bar{R}$ can be made available as a climatological entity, and that $\bar{E}$ will ultimately prove not to have significant dependence on the type and form of forest structure, although it may vary from one broad climatic region to the next.

The Gash model is still in an early stage of development but preliminary tests are promising. With such a model, the Thom-Oliver-Gash equation, equation (71), could prove a useful predictive formula of λE_T for tall vegetation, while equation (67) might well be useful in this context for short vegetation, given the relevant empirical value for m. Thom and Oliver suggest m = 1.4, partly as a value generally consistent with typical values of r_{sd}, but more particularly because this is the value required to ensure that the use of equations (67) and (68) yields the same annual estimate of λE_{RC} as the original Penman equation in the climate of south-east England where the Penman formula was calibrated.

(d) Energy Balance Models

In much the same way that the analysis of Shuttleworth (1976) provides the missing physical explanation of the link between simulation models and single source models, the analysis of Thom and Oliver (1977) provides the missing physical explanation of the link between single source and simpler energy balance models, of which the Penman equation is the original and typical example. This equation takes the form

$$\lambda E_{RC} = \frac{\Delta Q_n + \gamma f(u)(e_s - e)}{\Delta + \gamma} \qquad (74)$$

and provides an estimate of potential evaporation, λE_o, if Q_N is a measurement of net energy input relevant to a free water surface; or reference crop evaporation, λE_{RC}, if Q_N is relevant to short, green vegetation (Rijtema, 1965). Thom and Oliver (1977) demonstrate that the wind function

$$f(u) = 0.26\ (1 + U/100) \qquad (75)$$

used in equation (74) is implicitly serving two purposes namely,

(i) It contains a reasonable average description of the effect of thermal stratification on a rigorous formula for aerodynamic resistance with an assumed value of sensible heat flux of $H = 50\ W\ m^{-2}$. This value is not untypical of the conditions under which the Penman equation was derived.

(ii) When compared to physically rigorous formulae of aerodynamic resistance in neutral conditions, the implicit (small) value of z_o it contains is such as to compensate for the absence of an effective

surface resistance in the denominator, for crops in which $r_s/r_a \sim 1.4$ - at least in the meteorological region used to calibrate the original formulation.

The presence of this physical justification brings increased confidence to the use of equation (74) as a basis for estimating λE_{RC}, which usage is already widespread on the basis of past practical success. The use of equation (74) is recommended as the best available method of estimating E_{RC} and should be attempted as the primary means of obtaining this standard evaporation rate whenever the relevant data are available or can be obtained. The necessary measurements are those of dry bulb temperature, wet bulb temperature, wind run and net radiation. The first three are fairly common observations but, unfortunately, net radiation is not always available as a measurement and has to be estimated from solar radiation measurements, or measurements of sunshine hours. Doorenbos and Pruitt (1977) provide an excellent and simple description of how the Penman equation can be applied in practice: this is not repeated here.

Equation (74) (or minor modifications of it, not discussed here) is the simplest physically based equation which can be used to estimate λE_{RC}. Other meteorologically based classes of estimation techniques, lower in the hierarchy of Table 1, implicitly require additional empirical assumptions about the correlation of the meteorological parameters appearing in the Penman equation. Such correlations do exist, and do have some intuitive physical basis, but there must always remain doubt as to the universality of such empiricism in moving from one climatic region to the next. Equation (74) contains empirical relationships which should be tested in each application if possible, but the empirical relationships outlined in the following sections involve a higher level of empiricism in their estimation of E_{RC}, and should be used with care unless they can be calibrated against actual measurements of reference crop evaporation in local conditions.

(e) Radiation Models

Of the remaining classes of meteorological estimation techniques, the class most likely to preserve some of the assumed universality of the Penman equation is that with empirical equations which relate reference crop evaporation to radiation, or more correctly to a combination of temperature and radiation. The equation we take as representative is that of Priestley and Taylor (1971), an equation with increasingly popular usage, which has the form

$$\lambda E_{RC} = \alpha \frac{\Delta}{\Delta + \gamma} Q_N \qquad (76)$$

where α is an empirical constant in the order of 1.26.

The conceptual basis for equations of this type is the presence of an empirical relationship between the two terms in the numerator of equation (74), *viz*

$$(e_s - e) = (\text{constant}) \frac{\Delta Q_n}{f(u)} \qquad (77)$$

Appropriate relationships of this type are observed and are not unreasonable on physical grounds. Increased radiation will indeed tend to generate increased temperature and increased humidity in the atmosphere, as a result of the in-going surface fluxes of sensible and latent heat. The size of the increase will probably be reduced at higher windspeed because the volume of air affected by the surface fluxes is increased. (Note the presence of f(u) in the denominator of equation (77)). Since the relationship between saturated vapour pressure and temperature is non-linear, a joint increase in both temperature and humidity will very probably give rise to the increase in vapour pressure deficit $(e_s - e)$ required by this equation. Moreover, the rate of increase in the deficit caused in this way is greater at higher temperatures, quite possibly in rough proportion to Δ, which appears in the numerator of equation (77).

It should be emphasized immediately that the above argument is really no more than a rationalization, and should not be taken as evidence for equation (76). At the same time it is worth noting Penman's (1956) observation that, for European conditions at least, the first term in the Penman equation is commonly four to five times as great as the second. This observation is consistent with the value $\alpha = 1.26$ and might be taken as some justification for allowing a higher level of empiricism in the less significant second term. The justification for using equation (76) in preference to equation (74) is that it does not require the measurement of humidity, which can be difficult. Such an equation would be simpler to use in conjunction with satellite data to calculate reference crop evaporation for large, remote areas where few data are currently available.

It should be made explicit that a test of the merit of the Priestley-Taylor equation in reference to the Penman equation, is equivalent to testing the hypothesis.

$$\gamma f(u) \ (e_s - e) = (\alpha - 1) \ \Delta Q_n \qquad (78)$$

Such a test does not require the measurement of actual or reference crop evaporation and could be carried out on a worldwide basis wherever the required meteorological data are to hand. Bearing in mind the possibility of using equation (76) with satellite data, such an exercise could well be worthwhile as a necessary preliminary to the large scale estimation of reference crop evaporation (using equation 76)) for use in developing countries.

Although equation (76) might provide a useful estimate of λE_{RC}; in common with the Penman equation, it should not be considered as a means of estimating λE_T, except for short green crops. Recent work by Shuttleworth and Calder (1979) has demonstrated the inapplicability of

the Priestly-Taylor equation in estimating potential evapotranspiration for forest vegetation. This work does however suggest that the equation might be of some use in estimating *transpiration* from a forest stand, providing the constant α is altered to 0.72. It is then possible to create an equation, with a basis very similar to that of the Thom-Oliver-Gash model, which takes the form

$$\lambda E_T = 0.72 \frac{\Delta}{\Delta + \gamma} Q_n + I\ (1 - c) \qquad (79)$$

This equation is a useful estimate of potential evaporation for tall crops in the UK, but remains untested elsewhere.

(f) Humidity Methods

It is of course possible to express an assumed relationship between the two terms in the Penman equation in the opposite form, and create empirical relationships between reference crop evaporation and vapour pressure deficit. Such expressions, which are modifications of the Dalton equation, have the general form

$$\lambda E_{RC} = f'(u)\ (e_s - e) \qquad (80)$$

in which f' is an empirical expression derived for the location of interest. Clearly the universality of such an equation is likely to be less than those involving radiation, since in this case the smaller of the two terms in equation (74) is being used to estimate the greater. Moreover humidity data are more difficult to obtain than radiation data; and when they are available in conjunction with windspeed, it is often possible to use the Penman equation as a superior estimate with estimates or measurement of net radiation.

(g) Temperature Methods

Many empirical formulae exist relating λE_{RC} to temperature: in keeping with the philosophy of this report we select one example, the Blaney-Criddle (1950) equation, on no other grounds than that this is widely used. The equation is designed to provide daily estimates of evaporation averaged over a month and takes the form

$$\lambda E_{RC} = C_u\ d_1\ (0.46T + 8) \qquad (81)$$

where C_u is the so called 'consumptive use' factor, reported in engineering literature but best determined locally.

d_1 is the fraction of daylight hours occurring in the month

and T is the temperature in ^{o}C (Note that the numerical constants in this equation merely reflect the fact that empirical constants have usually been derived with temperature in ^{o}F).

Clearly the physical basis for this equation is merely that both the 'radiation' and 'aerodynamic' terms in the Penman equation are likely to have some, albeit ill defined, relationship with temperature. Since the radiation term is generally the larger of the two, it is the correlation between radiation and temperature which is more important. There is a lag in the yearly temperature cycle with respect to the yearly radiation cycle, and the monthly 'consumptive use' factor often, unwittingly, includes some allowance for this thermal lag.

The only real justification for using estimation methods of this type is that a *prediction* of evaporation is required on the basis of *existing* data, and temperature is the only measurement available. In this situation the work of Doorenbos and Pruitt (1977) provides a useful, practical reference. The present ready availability of net radiometers, which allow a Priestley-Taylor estimate, and/or evaporation pans makes the use of such equations in an 'estimation measurement' mode an unnecessary risk.

(b) 'Direct' Methods

One of the earliest methods of estimating evaporation does not fall naturally into the hierarchal scheme of Table 1. This consists of relating potential evaporation, λE_o, in the first instance, to the measured evaporation from pans of water, or from atmometers (small, wetted, porous surfaces). The measurement corresponds to the joint determination of a complex, but not irrelevant, combination of meteorological factors. Strictly this combination is only relevant to the evaporation measurement actually made, but, over appropriate time periods, there is indeed a high correlation with the standard rates of evaporation, suggesting that the combination is not too different to that relevant to other evaporating surfaces.

(i) *Atmometers*

Many types of atmometer exist (see for instance Ganopadhyaya *et al*, 1966), of which the Piche atmometer is an example. Here the evaporation rate is measured from flat, horizontal disks of wetted blotting paper, 3.2 cm in diameter, with both sides exposed to the air. The energy balance of such an atmometer differs greatly from that for a free water surface and from natural vegetation. The energy for evaporation comes from radiation and convective transfer with the porous surface, and by heat conduction through the water supply system from other exposed surfaces. They are placed some height above the ground, usually well above the vegetation, and details of siting, and the conditions of surrounds and fetch, can seriously affect the response. None-the-less a consistently sited and well maintained atmometer *can* yield a reasonably good and consistent calibration against λE_o (or even λE_{RC}). However the problems of maintenance, siting and fetch have such large effects that universality in such calibration constants is largely illusory, and evaporation pans are the preferred 'direct' measurement method.

(ii) *Evaporation Pans*

In their description of evaporation pans more than ten years ago, Gangapadhyaya *et al* (1966) list 27 examples and suggest that their tabulation 'is undoubtedly far from complete'. It could be argued that in this situation the major criticism of pan usage is the diversity of its application rather than validity of the measurement it produces. This diversity is unhealthy in that it distracts attention away from the primary issue, whether an evaporation pan does or does not provide a number which is (or can be easily related to) potential evaporation, into the side issue of intercalibration of pan types.

It is clear that the energy exchange of all pans will differ from that of a reference crop. Firstly the energy storage (particularly of deep pans) is greater than that of vegetation so that the surface temperature of the water tends to be lower during the day, when most evaporation occurs, and higher at night. If the tank is elevated above the surface, additional radiation exchange at the sides, and sensible heat exchange at the sides and surface, can also give rise to differences. Bearing in mind the additional levels of empiricism necessary to turn a measurement of pan evaporation into an estimate of actual evaporation, enough work has already been done in assessing the relative merits of pan types, and in the detailed description of their energy exchange. A single design, the 'Class A', has now been selected as a world wide standard for use in all new applications.

It is hoped that ultimately, simple, reliable instrumentation, capable of measuring actual evaporation, will be developed, and become cheap enough to remove the need for this usage. Meanwhile, evaporation pans will certainly continue to be used as a low technology replacement for the present alternative of an 'estimation measurement' based on (say) the Penman equation. At the same time the widespread availability of existing pan data, albeit from diverse designs, means that a great many predictive estimation applications will continue to use evaporation pans as a primary source for a considerable time. It is therefore fortunate that there is often a good correlation between the 'standard rates' of evaporation, λE_o and λE_{RC}, and measured evaporation from pans, and that a great deal of empirical data are available describing this correlation, at least for the more common types of pan. The reader is referred to the recent work of Doorenbos and Pruitt (1977) for a useful, practical description of how pan estimation of E_{RC} might be carried out for two of the most common types, namely the 'Class A' pan and the 'Colorado sunken pan'. The high level of empiricism implicit in their description should be recognised as a real and permanent influence on the reliability of pan estimation techniques.

3. COMMENTS ON RELATING 'ACTUAL' TO 'POTENTIAL' EVAPORATION

For practical applications, it is actual evaporation which is most often required but a great many of the techniques used in estimating evaporation, (including all of those in most common use) usually provide one of the 'standard' rates, generally λE_{RC}. The conventional practice is then to multiply this standard rate by an additional factor, *ie*

$$\lambda E = K_c \lambda E_{RC} \tag{82}$$

The factor K_c in this equation is called the crop coefficient and, from a study of equation (57) it can be seen to be a measure of the term

$$K = \frac{\Delta + \gamma}{\Delta + \gamma (1 + r_{s/r_a})} \tag{83}$$

where r_a and r_s are the 'effective' aerodynamic resistance and surface resistance of the crop. Such a term is of necessity extremely complex. It will contain some dependence on meteorological parameters; both explicitly, as temperature in Δ, as windspeed in r_a, and as rainfall in r_s (indirectly in the amount of time the canopy is wet or partially wet). It will also contain dependence on the vegetation structure, through r_a, and its stomatal behaviour (which is the primary source of any variation in r_s in dry conditions). The stomatal control itself might also be related to current or past meteorological parameters. Clearly the use of such a complex parameter should only be made with a proper realization of the risks involved. Although the basis and definition of the crop coefficient is essentially scientific, its complexity is such that its application is an art.

In practice a great many applications are concerned with the implementation of irrigation for agricultural crops. The objective is usually to supply water which is adequate, in that it does not limit growth, but not excessive, so that the soil surface is not wet. Such conditions are essentially those used to specify potential evapotranspiration (except there is some implicit but poorly defined suggestion of an additional restriction to low rainfall conditions). In this case an entity, which we will call a 'potential crop coefficient', K_{co}, might be relevant, and is defined from the equation

$$\lambda E_T = K_{co} \lambda E_{RC} \tag{84}$$

It is quite possible that K_{co} could be less variable than K_c in moving from one location to the next, since r_s is likely to be a purer measure of stomatal resistance, which in turn is probably less variable since the soil moisture deficit remains small. It will of course still have some local meteorological dependence through Δ and r_a.

Considerable work has been done on measuring the potential crop coefficient, K_{co}, as a function of time for different crops. As might be expected for agricultural crops, with which it is often used, there is a pronounced seasonal variation of the type illustrated schematically in Figure 9. Tables of likely values of K_{co} exist in the literature for certain parts of the world (*eg* USA and Israel), which may be used to provide an estimate of E_T, using equation (84), for different stages in the development of a great many irrigated annual crops. (See in particular pages 37-44 of Doorenbos and Pruitt, 1977). Irrigated perennial and fruit crops do not have such a marked variation in K_{co}, so that tables summarizing empirical estimates of the potential crop coefficient are more usually presented as a function of the time of year (*eg* pages 45-52, *ibid*).

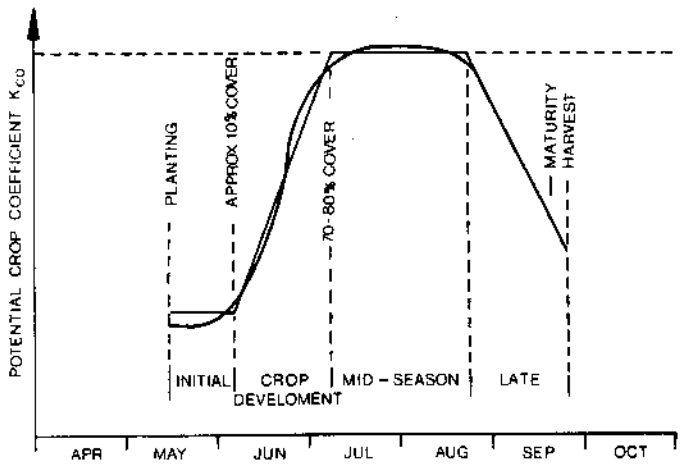

FIGURE 9

THE TYPICAL VARIATION IN POTENTIAL CROP COEFFICIENT, K_{co}, MEASURED FOR AN AGRICULTURAL CROP IN THE NORTHERN HEMISPHERE

If estimates of evaporation are required for non irrigated crops, the water status of the soil can become important through surface resistance: K_c becomes the relevant crop coefficient in such situations. This problem is often compounded by the fact that such estimates are very commonly required in water resource estimation applications. This usually implies the presence of some significant rainfall, and a consequent confused definition of the effective surface resistance in terms of both plant and meteorological parameters. In moist climates where rainfall occurs fairly frequently throughout most of the year, an average time-dependent empirical description of K_c (and by implication of r_s) is probably the best that can be attempted. Here the situation is mitigated by the fact that for a great many short crops the actual evaporation rate is commonly quite close to reference crop evaporation most of the time ($K_c \sim 1 \pm 15\%$); a result not unrelated to the 'cancellation effect' present in the Penman equation and described in the relevant section of this chapter. Tall crops (forests) remain the primary exceptions in such climates because of the real possibility of significant advective enhancement in the evaporation rate (Shuttleworth and Calder, 1979).

In climates with large seasonal variations in rainfall, evaporation rates in the rainy season will again probably be close to λE_{RC} for short crops, but might exceed λE_{RC} for tall crops, depending on details of the interception mechanism. In the dry season significant fall in K_c will usually be observed as the surface resistance responds to the increasing soil moisture deficit. Studies have been made of the variation in K_c in response to decreasing soil water content: the results can of course differ considerably in detail, as might be expected for an empirical parameter of this type, but many workers are in agreement that the over all behaviour in drying cycle follows the general pattern illustrated in Figure 10. The qualitative behaviour seems to be similar for both crops and soil: during the first fairly constant stage K_c remains quite close to its initial value until a 'wilting point' is reached when K_c for the crop (or soil) begins to decrease in response to decreasing soil water content. In conditions of prolonged drought the crop (if present) begins to die and the evaporation rate is no longer controlled by meteorological conditions, but by soil characteristics, especially hydraulic conductivity.

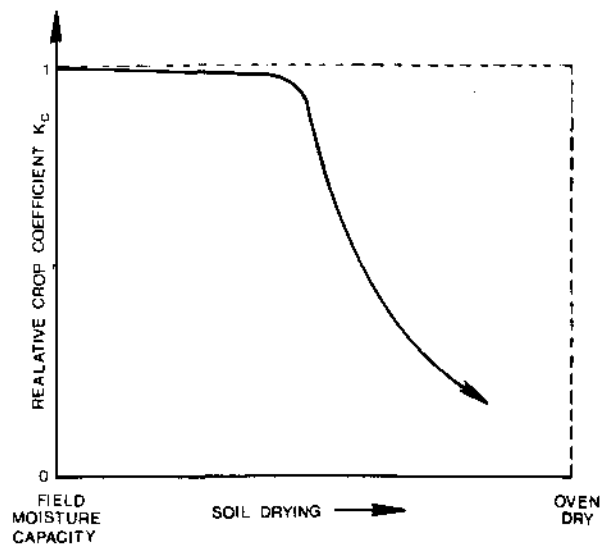

FIGURE 10

SCHEMATIC DIAGRAM OF THE TYPICAL BEHAVIOUR OF THE CROP COEFFICIENT AS A FUNCTION OF TIME DURING A DRYING CYCLE

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